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Bedrock Geologic Map of the Crown Point Quadrangle, Essex County, New York, and Addison County, Vermont



Pamphlet to accompany
Scientific Investigations Map 3491

Cover: View looking south into the Champlain Valley from the northern border of the Crown Point quadrangle. Outcrop is Lyon Mountain Granite Gneiss. Bulwagga Mountain is visible on the right (west) behind the lead author, Lake Champlain is visible in the center distance, and the Green Mountains are visible on the horizon to the left (southeast). Photograph by Gregory J. Walsh, U.S. Geological Survey.

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By Gregory J. Walsh, Randall C. Orndorff, and Ryan J. McAleer

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**U.S. Department of the Interior
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Contents

Acknowledgments	v
Abstract	1
Introduction.....	1
Stratigraphy	3
Mesoproterozoic Lithostratigraphy	3
Cambrian-Ordovician Stratigraphy	6
Gamma Radiation Measurements	9
Structural Geology.....	11
Ductile Structures.....	11
Brittle Structures.....	15
Veins.....	15
Joints.....	16
Faults.....	16
Champlain Valley Faults	16
Tectonics and Metamorphism	20
Economic Geology.....	21
Iron Mines.....	21
Graphite	22
Feldspar	22
Paleozoic Rocks	22
Phosphate	22
References Cited.....	22
Appendix 1. Representative Photographs of Map Units From the Crown Point Quadrangle	31

Figures

1. Simplified geologic map of the Adirondacks, upstate New York, showing the location of the Crown Point quadrangle	on map sheet
2. Photograph of the Great Unconformity.....	2
3. Photographs of the Lyon Mountain Granite Gneiss.....	4
4. Photographs of pegmatite and stereonet and rose diagram showing the orientations of measured pegmatite dikes.....	5
5. Photographs of a joint canyon on the east side of Buck Mountain with remnant fragments of a mafic dike exposed on a cliff face of biotite gneiss.....	10
6. Plane-light photomicrograph showing a cluster of three yellow-orange allanite crystals in a microcline granite sample of the Lyon Mountain Granite Gneiss	11
7. Stereonets and photographs showing the D2 structures.....	12
8. Stereonets and photographs showing F3 folds and related structures	13
9. Structural diagrams and photographs of boudinage or F4 folds.....	14
10. Structural diagram and photographs of D4 shear bands or shear zones.....	15
11. Structural diagrams showing the orientation of Ediacaran mafic dikes.....	17
12. Stereonet and photograph showing veins	17

13. Stereonet and rose diagram showing the orientations of measured joints	18
14. SEM images, powder X-ray diffraction spectrum, and photograph of a fracture- and fault-filling serpentine mineral from a marble sample collected from the Buck Mountain Pond graphite mine on Mine Hill	19
15. Stereonet showing the orientations of measured brittle faults	20
16. Map of the Crown Point quadrangle, Essex County, New York, and Addison County, Vermont, showing abandoned mines and quarries	on map sheet
17. Photograph of the Crown Point iron furnaces in about 1901	on map sheet
18. Photograph of magnetite ore seams at the Kent mine in host leucogranite gneiss	22
19. Stratigraphic column for Paleozoic rocks showing the approximate thickness of map units and conodont biozones	on map sheet

Tables

1. Summary of U-Pb zircon ages from the Crown Point quadrangle, Essex County, New York, and Addison County, Vermont	on map sheet
2. Analyses of conodont samples from the Crown Point quadrangle and adjoining Ticonderoga quadrangle, Essex County, New York, and Addison County, Vermont	7
3. Quantitative gamma radiation results on map units from the Crown Point quadrangle	9
4. Mines and quarries in the Crown Point quadrangle, Essex County, New York, and Addison County, Vermont	on map sheet

Conversion Factors

International System of Units to U.S. customary units

Multiply	By	To obtain
Length		
centimeter (cm)	0.3937	inch (in.)
meter (m)	3.281	foot (ft)
kilometer (km)	0.6214	mile (mi)
meter (m)	1.094	yard (yd)

U.S. customary units to International System of Units

Multiply	By	To obtain
Length		
mile (mi)	1.609	kilometer (km)

Datum

Horizontal coordinate information is referenced to the North American Datum of 1983.

Vertical coordinate information is referenced to the National Geodetic Vertical Datum of 1929.

Elevation, as used in this report, refers to distance above the vertical datum.

Abbreviations

AMCG	anorthosite-mangerite-charnockite-granite suite
Ga	giga-annum (billion years before present)
GIS	geographic information system
GPS	global positioning system
K-Ar	potassium-argon
km	kilometer
lidar	light detection and ranging
LMG	Lyon Mountain Granite Gneiss
m	meter
Ma	mega-annum (million years before present)
mi	mile
MRDS	Mineral Resources Data System
REE	rare earth element
SHRIMP	sensitive high resolution ion microprobe
U-Pb	uranium-lead
μ R/h	microrems per hour

Bedrock Geologic Map of the Crown Point Quadrangle, Essex County, New York, and Addison County, Vermont

Gregory J. Walsh, Randall C. Orndorff, and Ryan J. McAleer

Abstract

The bedrock geology of the 7.5-minute Crown Point quadrangle consists of deformed and metamorphosed Mesoproterozoic gneisses of the Adirondack Highlands unconformably overlain by weakly deformed lower Paleozoic sedimentary rocks of the Champlain Valley. The Mesoproterozoic rocks occur on the eastern edge of the Adirondack Highlands and represent an extension of the Grenville Province of Laurentia. Granulite facies Mesoproterozoic paragneiss, marble, and amphibolite hosted the emplacement of granitic orthogneiss at approximately 1.18–1.15 giga-annum (Ga, billion years before present). The earliest of four phases of deformation (D1) is characterized by gneissosity, rarely preserved F1 isoclinal folds, and migmatite in the host rocks. Subsequent D2 deformation produced a composite penetrative gneissosity, migmatite, and isoclinal F2 folds. Towards the end of D2, felsic magmatism (including the regionally extensive Lyon Mountain Granite Gneiss, abbreviated “LMG”) spread by penetrative migration as semiconcordant alkali feldspar granite sheets subparallel to S2 into previously deformed lithologies. The LMG crystallized at approximately 1.15 Ga and displays synkinematic F2 folds thus constraining the time of D2 deformation. Exhumation during D3 produced F3 folds exhibited in regional domes and basins, such as the Keeney Mountain synform, local reactivation of the S2 foliation, partial melting, metamorphism, metasomatism, iron ore remobilization, and intrusion of magnetite-bearing pegmatite both as layer-parallel sills and crosscutting dikes. D4 created NE- and NW-trending boudinage, local high-grade ductile shear zones, and crosscutting granitic pegmatite dikes. Kilometer (km)-scale lineaments readily observed in lidar data are Ediacaran mafic dikes and Phanerozoic brittle faults. The Paleozoic rocks are part of the Early Cambrian to Late Ordovician great American carbonate bank on the ancient margin of Laurentia. Cambrian-Ordovician stratigraphy records an approximately 1-km-thick section and a transition from synrift clastics to passive margin peritidal carbonate buildups to gradually deeper water subtidal to shelf carbonates during foreland basin development associated with the Taconic orogeny. The Paleozoic rocks are weakly folded and block faulted. Large areas of the Champlain Valley are covered by undifferentiated glacial deposits, some of which contain

mapped landslides. The map also shows waste rock piles and tailings from historical mining operations and large areas of artificial fill.

This study was undertaken to improve our understanding of the bedrock geology in the Adirondack Highlands, establish a modern framework for 1:24,000-scale bedrock geologic mapping in the Adirondacks, provide a context for historical iron mines in the eastern Adirondacks, and update the stratigraphy of the Champlain Valley in New York and Vermont. This Scientific Investigations Map of the Crown Point 7.5-minute quadrangle consists of a map sheet, an explanatory pamphlet, and a geographic information system database that includes bedrock geologic units, faults, outcrops, and structural geologic information. The map sheet includes a bedrock geologic map, a correlation of map units, a description of map units, an explanation of map symbols, three cross sections, and a simplified surficial geologic map that includes lidar percent slope. The explanatory pamphlet includes a discussion of the geology.

Introduction

The bedrock geology of the 7.5-minute Crown Point quadrangle consists of deformed and metamorphosed Mesoproterozoic gneisses of the Adirondack Highlands unconformably overlain by weakly deformed lower Paleozoic sedimentary rocks of the Champlain Valley (see bedrock geologic map and fig. 1 on map sheet). Unconsolidated Quaternary glacial sediments overly the bedrock and conceal much of the bedrock geology in the eastern part of the map (see simplified surficial geologic map on map sheet). The Mesoproterozoic rocks occur on the eastern edge of the Adirondack Highlands and represent an extension of the Grenville Province of Laurentia (fig. 1) (McLelland and others, 2013). The Paleozoic rocks are part of the Early Cambrian to Late Ordovician great American carbonate bank on the ancient margin of Laurentia (Landing, 2012), and their deposition on the Mesoproterozoic rocks created the Great Unconformity (fig. 2). Large areas of the Champlain Valley are covered in glacial deposits. On this map, the surficial geologic deposits are undifferentiated. Rayburn (2004) and De Simone and others (2008) discuss the surficial geology, which has yet to be mapped in detail in this quadrangle; Barker (1916) presented a preliminary map and first analysis of the surficial geology in the area. Small landslides

2 Bedrock Geologic Map of the Crown Point Quadrangle, Essex County, New York, and Addison County, Vermont

were also mapped during field work, but not studied. Areas of Holocene waste rock piles and tailings from mining are also mapped, as are large areas of artificial fill.

Prior to this study, the only published geologic map for the New York part of the Crown Point quadrangle was the 1:250,000-scale State geologic map of New York (Isachsen and Fisher, 1970; Rickard and others, 1970). In New York, prior published 1:62,500-scale mapping by Ogilvie (1905) exists for the adjacent Paradox Lake 15-minute quadrangle to the west. In Vermont, mapping at 1:62,500-scale by Welby (1961) included the Paleozoic geology to the eastern shore of Lake Champlain. Unpublished manuscript text and maps of 1:24,000-scale mapping by Walton (1966a, b) and 1:20,000-scale mapping by Rodgers (1950) were provided by the New York State Geological Survey and provided a foundation for our new mapping. Walton and de Waard (1963) summarized their mapping in a small-scale synthesis of the eastern Adirondacks, and it was largely Matt Walton's work in the area that provided the source for the geology depicted in this area of the Adirondack sheet of the State geologic map of New York (Isachsen and Fisher, 1970). The only published synthesis of unpublished quadrangle-scale mapping in the immediate area before this study is the work of Walton and de Waard (1963) and de Waard and Walton (1967). Their primary conclusion was a hypothesis that the paragneisses in the Adirondack Highlands represented a supracrustal sequence of rocks that post-dated the orthogneisses found in the area. This hypothesis disagreed with the seminal Adirondack work by Balk (1931) and Buddington (1939) who recognized that the igneous rocks intruded into a series of marble, quartzite, amphibolite, and paragneiss. Subsequent advances in Adirondack geochronology showed that the supracrustal hypothesis was incorrect, and the State geologic

map of New York (Isachsen and Fisher, 1970) reestablished the relative age of map units in the Adirondacks. Walton and de Waard (1963) recognized the dome and basin architecture in the eastern Adirondacks, and their distribution of major structures and map units is generally valid and useful for guiding ongoing and future geologic mapping in the area.

This study was undertaken to improve our understanding of the bedrock geology in the Adirondack Highlands, establish a modern framework for 1:24,000-scale geologic mapping in the Adirondacks, provide a context for historical iron mines in the eastern Adirondacks, and update the stratigraphy of the Champlain Valley.

The geologic mapping was conducted using recently acquired light detection and ranging (lidar) topographic data. Base maps included lidar percent-slope maps, high resolution leaf-off satellite imagery, 1:24,000-scale topographic maps, and recently acquired geophysical data including airborne magnetic and radiometric images (Shah, 2016; Shah and others, 2019, 2021). Digital data capture during mapping by Walsh was conducted with an iPad mini 4 running the Midland Valley Fieldmove app (v. 1.3). Data capture by Orndorff was conducted with an iPad mini 2 running the app IGIS HD. Locational information was obtained with a bluetooth enabled GPS receiver, either Garmin GLO or Bad Elf GPS Pro+ paired to the iPad. Field data were exported to Esri ArcGIS and final cartography was completed in Adobe Illustrator. A geographic information system (GIS) database accompanies the bedrock geologic map. Representative photographs of map units from the Crown Point quadrangle were taken during fieldwork and are presented in appendix 1.



Figure 2. Photograph of the Great Unconformity (at point of arrows). The Potsdam Sandstone rests unconformably above the migmatitic biotite gneiss member (Ybg) of the Grenville Complex on Putnam Creek near Crown Point Center. For scale, the log in the plunge pool is about 30 centimeters in diameter. Photograph by Gregory J. Walsh, U.S. Geological Survey.

Stratigraphy

Mesoproterozoic Lithostratigraphy

Paragneiss units of the Grenville Complex (Brown, 1979, 1988, 1989; Brown and Ayuso, 1985) are the oldest rocks in the quadrangle and the host rock for several generations of igneous rock. The Grenville Complex has alternatively been called the “Grenville supergroup” (Wynne-Edwards, 1972; Davidson, 1998; Chiarenzelli and others, 2015). The Grenville Complex is likely correlative with parts of the Mount Holly Complex in Vermont (Ratcliffe and others, 2011). The age range is not well constrained, but the complex is likely correlative, in part, with rocks in the Adirondack Lowlands, which range in age from about 1,284 to 1,258 mega-annum (Ma, million years before present) (Chiarenzelli and others, 2015). In the adjacent Eagle Lake quadrangle (Regan and others, 2015), the paragneiss units predate the intrusion of the granitic augen gneiss (Yggn) which yielded an age of $1,185 \pm 11$ Ma (Regan and others, 2019a). Importantly, the dated granitic augen gneiss to the west is contiguous with the Yggn unit mapped around Keeney Mountain in the Crown Point quadrangle, and thus provides an upper age limit for the paragneiss units in the region. Regionally, the paragneiss units are the host for widespread igneous rocks of the anorthosite-mangerite-charnockite-granite (AMCG) suite, which underly most of the Adirondack Highlands (Hamilton and others, 2004; McLelland and others, 2013). The anorthosite-mangerite-charnockite-granite (AMCG) suite has an age range of about 1,160 to 1,145 Ma (McLelland and others, 2013; Aleinikoff and others, 2021). Anorthosite, mangerite, and charnockite do not occur in the Crown Point quadrangle but are found in the Eagle Lake quadrangle to the west (Regan and others, 2015, 2019a). The most abundant paragneiss unit in the Crown Point quadrangle is the migmatitic biotite gneiss member (Ybg) of the Grenville Complex. Less abundant paragneiss units include the garnet-sillimanite gneiss (Ysi) and amphibolite gneiss (Ya) members of the Grenville Complex, and minor amounts of marble and calc-silicate rock (Ym and Ycs). Minor units that are slight variations of the migmatitic biotite gneiss member (Ybg) include the biotite gneiss with garnet (Ybgg) and the rusty weathering biotite-bearing paragneiss (Yrbg). In the region of the Keeney Mountain synform, interlayered amphibolite gneiss (Ya) and marble and calc-silicate gneiss (Ym) units are mapped to the south and east where they are largely obscured by the glacial overburden; there the two members are mapped as an undifferentiated unit (Yma). Exposures east of the Street Road delta and west of Route 9N/22 clearly show that the two units (Ya and Ym) are intricately interlayered.

Metagabbro (Ygb) and clinopyroxene-hornblende granitic orthogneiss (Yhg) are the next two youngest Mesoproterozoic rocks in the quadrangle, and they have limited exposure. Four metagabbro (Ygb) bodies are mapped in the northern and western parts of the quadrangle; one larger

body near Cold Spring Park, a smaller body west of Bush Road, and two small bodies on the western edge of the map south of Towner Hill Road. The largest metagabbro intruded both the Ybg and Yhg units in the northern part of the map, and the smaller bodies are exposed within the Ylg (leucogranite) and Ybg units where the contacts are not exposed. Ages from a metagabbro dike at Dresden Station, N.Y., in the eastern Adirondacks include $1,144 \pm 7$ Ma (McLelland and others, 1988; McLelland and others, 2004) and $1,134 \pm 3$ Ma (Walsh and others, 2016), but the metagabbro mapped in this quadrangle does not occur as dikes. Instead, the metagabbro bodies appear as intrusive plugs. Regional metagabbro bodies are described by Regan and others (2011). Clinopyroxene-hornblende granitic gneiss (Yhg) occurs in the northwestern part of the map as deformed sheet-like bodies that intruded paragneiss units Ybg and Ya. The rock has a noticeable deformation fabric and is invariably gneissic. Walton (1966a, b) mapped the rock as his hornblende granite gneiss unit (syg). The rock is compositionally similar to the Hawkeye Granite Gneiss, but the Hawkeye has fewer mafic minerals and texturally it possesses an inequigranular texture with feldspar augen, called “phacoidal texture” by Postel (1952, p. 9), that was not observed in the Crown Point quadrangle.

The most widespread orthogneiss in the quadrangle is the Lyon Mountain Granite Gneiss (Ylg and Ylgg map units, abbreviated LMG), which occupies almost half of the Mesoproterozoic exposure in the map. The name “Lyon Mountain Granite Gneiss” was formally adopted by J.B. Reeside, Chair of the U.S. Geological Survey’s Geologic Names Committee in March and again in May of 1950 (N. Stamm, U.S. Geological Survey, written communication, 2017) based on the definitions in submitted manuscripts that were subsequently published by Postel (1951, 1952). The rock was originally named the Lyon Mountain Granite by Miller (1919, 1926) for exposures in the hamlet of Lyon Mountain, located in the town of Dannemora in Clinton County, New York. The rock was called the “ore formation” granite by Kemp and Alling (1925, p. 51), and all workers recognized the occurrence of historically significant iron ore magnetite deposits within it or adjacent to it. Some confusion exists in the literature regarding proper nomenclature, and the rocks within the unit have also been called the Lyon Mountain Granitic Gneiss (McLelland and others, 2002), Lyon Mountain gneiss (Whitney and Olmsted, 1988, 1993; Foote and McLelland, 1995), Lyon Mountain Granite suite (Lupulescu and others, 2012; Chiarenzelli and others, 2017), or simply the Lyon Mountain Granite (Valley and others, 2009, 2010, 2011; Miller, 1919, 1926). Despite the nomenclature issue, most workers agree that it is the relatively youngest widespread granite exposed within the Adirondack Highlands (Buddington, 1939; Postel, 1951, 1952; Buddington and Leonard, 1962; McLelland and others, 2001, 2002; Selleck and others, 2005; Valley and others, 2009, 2010, 2011; Lupulescu and others, 2012; Chiarenzelli and others, 2017; Aleinikoff and others, 2021).

The Lyon Mountain Granite Gneiss is a ferroan A-type leucogranite (Chiarenzelli and others, 2017; Aleinikoff and others, 2021) that is locally hydrothermally altered (Valley and others, 2011). Alkali feldspar bearing phases locally show alteration to quartz-albite rock, most likely the result of fluid alteration, especially near magnetite seams (Foosse and McLelland, 1995; McLelland and others, 2002; Valley and others, 2011). Previous geochronology studies concluded that the Lyon Mountain Granite Gneiss intruded at about 1,050 Ma (McLelland and others, 2001; Selleck and others, 2005; Valley and others, 2009, 2010, 2011; Chiarenzelli and others, 2017). New interpretations of the U-Pb geochronology (Aleinikoff and Walsh, 2015; Walsh and others, 2016; Aleinikoff and others, 2021) have challenged the accepted paradigm, and report abundant evidence that shows the LMG is related to the “granite” part of the AMCG suite, with ages from nine samples that range from about 1,163–1,141 Ma. The Lyon Mountain Granite Gneiss may be similar to ferroan granites found in other granulite terranes which evolved from charnockitic melts to biotite or biotite-hornblende granites (Frost and Frost, 2008).

Several undifferentiated rock types occur in the mapped Ylg unit including micropertthite granite, microcline granite, quartz syenite, alkali feldspar granite (alaskite), syenogranite, monzogranite, and quartz-albite rock, all of which contain magnetite. Despite our higher resolution mapping (1:24,000 vs. 1:62,500), we agree with Postel (1951, 1952) that the different rock types cannot be mapped separately in the field, and subdivision therefore is “a microscopic rather than a megascopic task” (Postel, 1951). Since no one has of yet successfully mapped the different rock types (lithodemes) in the LMG, there is no justification for using the suggested term “Lyon Mountain Granite suite” (Chiarenzelli and others, 2017) according to the North American Stratigraphic Code (2005). The Lyon Mountain Granite Gneiss occurs as large sills, intrusive sheets, or slab-shaped plutons originally described as phacoliths by Buddington (1939). Layering is interpreted as mostly primary igneous layering formed as flow-banding during syn- to post-tectonic emplacement. Much of the fine layering does not appear to be due to subsolidus deformation processes as equigranular quartz and feldspar only rarely show grain size reduction and preferred grain shapes, and the rock has a general paucity of fabric forming minerals like biotite and hornblende, which are rarely aligned (fig. 3). Local tight to isoclinal folding (F2) of the banding suggests that the rock was emplaced at a late stage of the main foliation-forming event (D2).

The relatively youngest Mesoproterozoic igneous rocks include pegmatite (Yp) that occurs as mapped bodies in granitic, tabular semiconcordant to discordant sills or dikes, with smaller locally mapped dikes or sills shown on the map by strike and dip symbols. The semiconcordant bodies are foliation-parallel sills, and the crosscutting dikes occur as either tabular straight-walled dikes, pegmatite-filled ductile shear bands, axial planar segregations in F3 or F4 folds, and boudin necks related to D3 to D4 extensional fabrics (fig. 4, see Structural Geology section). Statistically, the measured tabular dikes show prominent N and NE trends (fig. 4E).

Pegmatite occurs in all the Mesoproterozoic rocks but is not mapped everywhere at the scale of the bedrock geologic map, and masses of irregularly shaped unmapped pegmatite occur throughout the area. Most of the mapped pegmatite is relatively young and crosscuts the dominant foliation in the rock, but relatively older pegmatite is also present in the migmatitic paragneiss units as evidenced by locally preserved isoclinally folded and highly foliated pegmatite segregations. Pegmatites in the Adirondack Highlands range in age from 1,222 to 949 Ma, but crosscutting pegmatites range from about 1,090 Ma to 949 Ma (Lupulescu and others, 2011). Valley and others (2011) report ages of 1,030 to 1,016 Ma from crosscutting pegmatites in the Lyon Mountain Granite Gneiss (Ylg). Two dated pegmatites (Yp) in the Crown Point quadrangle include localities at the Crown Point Spar Company quarry (Crown Point, 1,025±1 Ma), and 65 meters (m) south of Sugar Hill Road on the west side of Route 9N/22 (Sugar Hill, 1,048±14 Ma) (Lupulescu and others, 2011; M. Lupulescu, New York State Geological Survey, unpublished data, 2019) (table 1 on map



Figure 3. Photographs of the Lyon Mountain Granite Gneiss (Ylg). *A*, Hand sample shows the typical light-pink, equigranular, weakly foliated granitoid; and *B*, an outcrop photograph shows rare, strained quartz and feldspar in a late-D4 shear zone on the south side of Bulwagga Mountain; this subsolidus fabric is only locally observed near shear zones and is not widespread. Pencil in *A* and compass in *B* are shown for scale. Photographs by Gregory J. Walsh, U.S. Geological Survey.

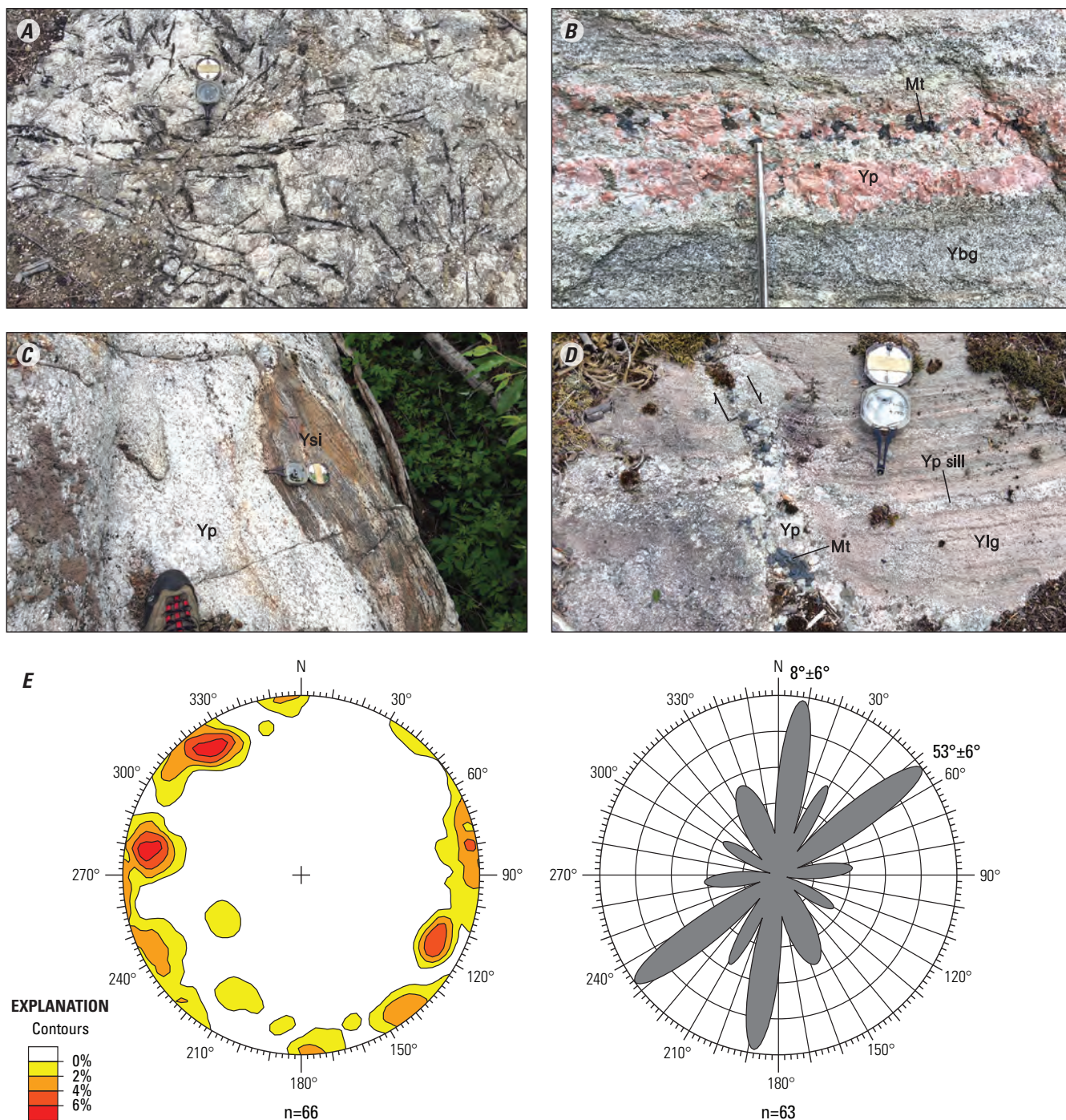


Figure 4. A, Photograph of pegmatite (Yp) showing a coarse-grained unfoliated variety with very coarse biotite. Compass is shown for scale. B, Photograph of a layer-parallel pegmatite (Yp) sill with coarse magnetite (Mt) in migmatitic biotite gneiss (Ybg). Note that the pegmatite is not foliated. Magnetic pick-up tool is shown for scale. C, Photograph of a compositionally banded pegmatite (Yp) intruded into previously deformed rusty gray garnet-sillimanite gneiss member (Ysi) of the Grenville Complex. Compass is shown for scale. D, Photograph of pegmatite (Yp) with coarse magnetite (Mt) in a shear band and as a sill within Lyon Mountain Granite Gneiss (Ylg). Arrows show direction of shearing. Compass is shown for scale. E, Stereonet (left) and rose diagram (right) showing the orientations of measured pegmatite dikes. The stereonet shows a lower-hemisphere equal area projection of poles and contoured poles to dikes. The rose diagram shows a normalized subset of the data for dips $>59^\circ$, and statistical peak trends are shown with 1 standard deviation error (for example, $8^\circ \pm 6^\circ$). The number of structural measurements in each dataset is indicated by "n" at the bottom of each diagram. Stereonet and rose diagram were plotted using the Structural Data Integrated System Analyser (DAISY 3, version 5.14a) software by Salvini and others (1999) and Salvini (2016). Photographs by Gregory J. Walsh, U.S. Geological Survey.

sheet). A third dated pegmatite (Yp) occurs as a 1-m-thick pink and white granite pegmatite dike with magnetite that intruded a hornblende-biotite paragneiss mapped within unit Ybg about 10 m south of the mined magnetite at the uphill (west) end of the Vineyard mine trench. The dike is in a right-lateral shear band that cuts the dominant foliation at a low angle; the location is shown by the pink strike-and-dip symbol on the bedrock geologic map. The dike of Yp exhibits crude layering of green alteration zones in the rock-containing abundant epidote; these zones are subparallel to the walls of the dike. Plagioclase is white and K-feldspar is pink. The pegmatite shows hematite alteration around magnetite. Plagioclase is highly saussuritized. Some late microcline could be related to late epidote with dissolution of plagioclase releasing anorthite content with calcium and aluminum forming Al-rich epidote (Helgadóttir and others, 2015). U-Pb zircon geochronology by the sensitive high resolution ion microprobe II (SHRIMP-II) yields ages of $1,034 \pm 5$ Ma on igneous cores and $1,024 \pm 5$ Ma on metamorphic rims (table 1; J.N. Aleinikoff, U.S. Geological Survey, unpublished data, 2020).

The Great Unconformity is a worldwide gap in the geological record between 100 million and 1 billion years long (DeLucia and others, 2017) that in most regions separates continental crystalline basement rock from much younger Cambrian shallow marine deposits (Yochelson, 2006; Peters and Gaines, 2012). In the Crown Point quadrangle, the Great Unconformity is marked by the Potsdam Sandstone overlying Mesoproterozoic igneous and metamorphic rocks (fig. 2) and is a gap of about 525 million years (Yp pegmatites, ~1,025 Ma, overlain by Potsdam Sandstone, ~500 Ma).

Cambrian-Ordovician Stratigraphy

There has been much disagreement on the Cambrian and Ordovician stratigraphy of the Champlain Valley in New York and Vermont dating to the late 1800s (Brainerd, 1891) and continuing into the 21st century (Landing, 2012). The variation in carbonate and clastic lithofacies deposited along the east side of the Adirondack massif during the early Paleozoic is the basis of this confusion. Fisher (1962) stated that the stratigraphy is “conflicting and almost unreconcilable.” For the Crown Point quadrangle, the authors carefully considered the history of stratigraphic names and examined outcrops to develop the map units for this part of the Champlain Valley. Although many references were consulted, stratigraphic descriptions from various works were most helpful (Rodgers, 1937; Oxley and Kay, 1959; Welby, 1961; Fisher, 1962; Fisher, 1968; Isachsen and Fisher 1970; Fisher, 1984; Washington and Chisick, 1988; Mango, 1997; Landing and others, 2003; Landing and Westrop, 2006; Landing and others, 2007; Ratcliffe and others, 2011; Landing, 2012; Dawson, 2015; Selleck and Mehrtens, 2015).

The Cambrian-Ordovician stratigraphy records a transition from synrift clastics to passive margin peritidal carbonate buildups (Potsdam Sandstone through Providence Island Dolomite) to gradually deeper water subtidal to shelf

carbonates during foreland basin development associated with the Taconic orogeny (Crown Point Limestone through Stony Point Formation) (Cady, 1969; Landing and Webster, 2018). This change in depositional setting is regional and can be seen in rocks throughout the Appalachians.

Paleozoic rocks in the Crown Point quadrangle range from Middle(?) Cambrian to Upper Ordovician and consist mostly of carbonate lithologies with clastic components. The oldest rocks deposited on the Great Unconformity are sandstones and conglomerates of the latest Middle(?) Cambrian Potsdam Sandstone (Єpt). The name Potsdam Sandstone has been used since it was introduced by Emmons (1838) and named for occurrences at Potsdam, St. Lawrence County, N.Y., 145 kilometers (km) (90 miles [mi]) northwest of Crown Point. Rodgers (1937) first mapped the Potsdam Sandstone in the Crown Point area. Although the Potsdam Sandstone has been considered Upper Cambrian, Landing (2012) reported it as uppermost Middle Cambrian and Upper Cambrian in the eastern and southern Champlain lowlands based on the trilobite *Crepicephalus* (Landing and Webster, 2018). The Potsdam Sandstone in the Crown Point quadrangle was mapped in stream valleys west of Crown Point Center where it overlies Mesoproterozoic gneiss and mapped along the western shore of Lake Champlain at Spar Mill Bay. The lower part of the formation is arkosic and contains pebble conglomerates and grades upwards to coarse- and medium-grained, greenish-gray sandstone, and records synrift fluvial and supratidal beach deposition (Fisher, 1968; Lowe and others, 2017). Selleck (1997) noted that the lower facies of the Potsdam Sandstone is greenish due to ubiquitous iron-rich chlorite. As the Potsdam was deposited on a high-relief surface of Mesoproterozoic rocks, its thickness is quite variable.

Rocks of the overlying Ticonderoga Formation (Єti) record the earliest deposition of carbonate rocks in the Paleozoic as platform dolostone with interbeds of detrital sandstone and sandy dolostone. The rocks of the Ticonderoga Formation were originally called the Galway Formation by Clarke (1910) for east-central New York with a type section designated by Fisher and Hanson (1951) at Galway, 113 km (70 mi) southwest of Crown Point. Welby (1959, 1961) named this interval the Ticonderoga dolostone and designated a type section near Mount Hope Cemetery and credited the name to John Rodgers who mapped the geology of the area. Isachsen and Fisher (1970) called the interval the Ticonderoga Formation. Landing and others (2003) reintroduced the name Galway and abandoned the term Ticonderoga. However, stratigraphic nomenclature that is still useful in its type area, should not be abandoned in that area (North American Stratigraphic Code, 2005; Owen, 2009). Since the type section is adjacent to the Crown Point quadrangle, the term Ticonderoga is used, herein.

In the Crown Point quadrangle, the Whitehall Formation (OЄw) overlies the Ticonderoga and records the dominance of carbonate deposition in the Late Cambrian. Rodgers (1937) named the Whitehall Formation for rocks above the Little Falls Dolomite and below the Cutting Dolomite with the type

section at Skene Mountain, Whitehall, N.Y., about 40 km (25 mi) south of the Crown Point quadrangle. However, the Little Falls Dolomite and Whitehall Formation have since been considered to include the same stratigraphic interval. Landing and others (2003) discussed abandonment of the Whitehall Formation as he considered it a “junior synonym” of the Little Falls. Since the Little Falls is named for rocks in Herkimer County, N.Y., and does not have a designated type section, the term “Whitehall” is used, herein. Although the Cambrian-Ordovician boundary has traditionally been placed in the upper part of the Whitehall Formation (Washington and Chisick, 1988), Landing and others (2007) stated that an unconformity exists at the top of the unit in the lower Champlain Valley and believed the Cambrian-Ordovician boundary occurs within the hiatus. However, conodonts (table 2) collected from the upper beds of the Whitehall Formation just south of the Crown Point quadrangle places this interval in the *Rossodus manitouensis* Biozone and confirms that the Cambrian-Ordovician boundary occurs in the Whitehall Formation.

The Cutting Dolomite (Ocu), which overlies the Whitehall Formation, contains quartz sand in its distinctive lower part as a constituent of crossbedded sandy dolostone. Cady (1945) named this interval the Cutting Dolomite for Cutting Hill in Shoreham, Addison County, Vt., 11 km (7 mi) east of the Crown Point quadrangle, with a type locality on the east dip slope of the hill. Landing (2003) and Landing and others (2003) abandoned the name Cutting Dolomite in favor of the “Tribes Hill Formation” first used by Ulrich and Cushing (1910) and named for Tribes Hill, Montgomery County, N.Y., about 129 km (80 mi) southeast of the quadrangle. The Tribes Hill Formation is a New York term that has not been previously mapped in Vermont. However, the Cutting Dolomite was mapped in Vermont by Ratcliffe and others (2011). Since the name Cutting Dolomite has been used locally in New York and Vermont and is lithologically different than the type rocks at Tribes Hill, the term “Cutting” is used herein. Samples of the Cutting Dolomite from the Crown Point and Ticonderoga quadrangles contain conodonts from the Lower Ordovician *Rossodus manitouensis* Biozone (table 2).

Table 2. Analyses of conodont samples from the Crown Point (CP) quadrangle and adjoining Ticonderoga (Ti) quadrangle, Essex County, New York, and Addison County, Vermont.

[Abbreviations: E, easting; m, meters; N, northing; NAD 83, North American Datum of 1983, zone 18N; UTM, Universal Transverse Mercator]

UTM coordinates (NAD 83)	Field sample number	Conodont fauna	Conodont biozone	Series (stage)	Conodont Alteration Index (CAI)	Formation (map unit)
626,052m E, 4,871,479m N	CP-3000	<i>Drepanoistodus</i> sp., <i>Glyptotonus quadriplicatus</i> , <i>Oepikodus communis</i>	<i>Oepikodus communis</i>	Lower Ordovician (Floian)	4.5–5	Fort Cassin Formation (Ofc)
627,943m E, 4,861,800m N	CP-3041	<i>Pseudobelodina manitouensis</i> , <i>Curtognathus</i> sp.	<i>Plectodina aculeata</i>	Upper Ordovician (Sandbian)	4.5–5	Orwell Limestone (Oo)
628,650m E, 4,859,279m N	CP-3047	<i>Drepanoistodus</i> sp., <i>Glyptotonus quadriplicatus</i>	<i>Oepikodus communis</i>	Lower Ordovician (Floian)	4.5–5	Fort Cassin Formation (Ofc)
629,219m E, 4,869,261m N	CP-3072	<i>Belodina compressa</i> , <i>Cahabagnathus</i> sp., <i>Plectodina aculeata</i> ?, <i>Pseudobelodina manitouensis</i>	<i>Belodina compressa</i>	Upper Ordovician (Sandbian)	4.5–5	Orwell Limestone (Oo)
626,627m E, 4,862,687m N	CP-3075	<i>Belodina compressa</i> , <i>Curtognathus</i> sp., <i>Panderodus</i> sp.	<i>Belodina compressa</i>	Upper Ordovician (Sandbian)	4.5–5	Orwell Limestone (Oo)
628,086m E, 4,855,671m N	Ti-3002	<i>Rossodus manitouensis</i> , <i>Loxodus bransoni</i> , <i>Loxodontatus bipinnatus</i>	<i>Rossodus manitouensis</i>	Lower Ordovician (Tremadocian)	4.5–5	Cutting Dolomite (Ocu)
626,626m E, 4,857,473m N	Ti-3113	<i>Rossodus manitouensis</i> , <i>Variabiloconus bassleri</i>	<i>Rossodus manitouensis</i>	Lower Ordovician (Tremadocian)	4.5–5	Whitehall Formation (Ocw)

Through most of the Champlain Valley, the Fort Ann (or Rochdale) Limestone overlies the Cutting Dolomite and is overlain by the Fort Cassin Formation. However, in the Crown Point quadrangle, no rocks attributable to the Fort Ann were mapped. Although this interval is mostly covered by glacial deposits, in areas where exposures of the Cutting and Fort Cassin Formations exist, sandy dolostone of the Cutting Dolomite gives way up-section to dolostone of the Fort Cassin Formation. Fisher (1984) mapped an unconformity between the Fort Ann Limestone and Fort Cassin Formation while Ratcliffe and others (2011) used the term “Bascom Formation” in Vermont for both units. Landing (2012) shows unconformities below and above the Fort Ann Limestone but uses the older name Rochdale Limestone instead of Fort Ann Limestone. Mango (1997) shows the Fort Cassin Formation directly overlying the Cutting Dolomite in the adjacent Orwell, Vt.-N.Y., quadrangle to the southeast of Crown Point. With the lack of limestone in this interval in the Crown Point quadrangle, and potential unconformities, we herein follow Mango (1997) and have not mapped the Fort Ann Limestone.

The Fort Cassin Formation (Ofc) was first used by Whitfield (1890) for interbedded dolostone and limestone below the Providence Island Dolomite for Fort Cassin, Addison County, Vt., just north-northeast of the Crown Point quadrangle. Washington and Chisick (1988) stated that the Fort Cassin Formation disappears in the western part of the Champlain Valley and is absent at Ticonderoga, N.Y., and they called rocks of this interval the Lemon Fair Formation. The Fort Cassin Formation was described as mostly limestone with subordinate dolostone. However, in the Crown Point quadrangle, it is dominantly dolostone in outcrop; the glacial surficial deposits may be masking the limestone. Although the rocks in the interval between the Cutting and Providence Island Dolomites within the Crown Point quadrangle are dominantly dolostone, the name Fort Cassin Formation is used herein as this name has been used in both Vermont and New York. The conodont *Oepikodus communis* was recovered from samples of the Fort Cassin Formation and places the formation in the upper part of the Lower Ordovician (table 2).

The Providence Island Dolomite (Opi) overlies the Fort Cassin Formation in the Crown Point quadrangle and a type section was described by Erwin (1957) for exposures on Providence Island, Vt., in Lake Champlain about 80 km (50 mi) north of Crown Point. This unit is generally a massive dolostone with thin limestone interbeds. The Lower-Middle Ordovician boundary occurs either at the base of the Providence Island or within the lower part of the formation. Samples from the Crown Point quadrangle have produced no conodonts, and biostratigraphic information is sparse, however, Landing and Westrop (2006) reported conodonts from the *Paraprioniodus costatus* interval indicating an early Middle Ordovician age for the formation.

The Knox unconformity exists at the top of the Providence Island Dolomite, which is overlain by the Day Point Formation, Crown Point Limestone (Ocp), and Valcour Limestone (Ov) to the north of Crown Point. The unconformity cuts out rocks of the Day Point Formation throughout the Crown Point quadrangle. Oxley and Kay (1959) noted that the Day Point Formation pinches out from north to south where it is absent at Crown Point. However, in the southeast part of the Crown Point quadrangle, high-calcium fine-grained limestone is in contact with dolostone of the Providence Island Dolomite. This suggests that the Knox unconformity cuts out the Crown Point and Valcour Limestones as well as the Day Point Formation. Fisher (1968) shows these units pinching out from the northern to southern Champlain Valley with the Orwell Limestone in contact with the Providence Island Dolomite in the south. This transition of erosion or nondeposition of the Crown Point and Valcour Limestones occurs in the Crown Point quadrangle where the Crown Point Limestone exists in the northeast part of the quadrangle and variably in the southeastern part of the quadrangle. Geologic mapping in the southeastern part of the quadrangle shows either an angular unconformity or high relief of the unconformity surface. The contact between the Providence Island Dolomite and the overlying limestone (Crown Point Limestone or Orwell Limestone) is sharp and represents the Knox unconformity (Landing, 2003; Landing and Webster, 2018), a regional disconformity that has been recognized throughout the Appalachians. In the southeastern part of the Crown Point quadrangle, the Orwell Limestone lies above the Providence Island Dolomite as conodont biostratigraphy shows that high-calcium limestone above the unconformity is within the Upper Ordovician *Belodina compressa* Biozone (table 2). The Crown Point Limestone was originally referenced by Cushing (1905) and described by Cady (1945). The bulk of the formation is thin- to medium-bedded limestone with argillaceous and silty laminations. The limestone of the upper part of the Crown Point Limestone is sandy containing quartz and feldspar grains. A common fossil within the Crown Point Limestone is the large gastropod *Maclurites*, which was observed in outcrops in the northeastern part of the quadrangle north of the town of Crown Point and along the western shore of Lake Champlain at Phelps Rock. Landing (2012) interpreted unconformities at both the base and top of the Crown Point Limestone with the Middle-Upper Ordovician boundary occurring within the formation.

The Valcour Limestone (Ov) and the Orwell Limestone (Oo) do not crop out well. Limestones assigned to these formations are best exposed at the Crown Point State Historic Site north of the quadrangle boundary in the Port Henry quadrangle in New York. The Valcour Limestone was first referenced by Cushing (1905) and described by Oxley and Kay (1959) who designated a reference section on Valcour Island in Lake Champlain near South Hero, Vt., about 80 km (50 mi) north of the quadrangle. Cady (1945) named the Orwell Limestone for Orwell, Addison County,

Vt., just southeast of Crown Point. A prominent bed of dolostone occurs near the contact between the Valcour and Orwell Limestones that has been placed in either formation by various workers. Welby (1961) describes the Valcour as being dolomitic, and Selleck and Mehrtens (2015) placed the dolostone in the Valcour Limestone at the Crown Point Historic Site section. However, Dawson (2015) placed the dolostone at the base of the Orwell Limestone. As the Valcour contains dolomitic limestones, this dolostone bed is placed at the top of the Valcour Limestone, herein.

The Glens Falls Limestone (**Ogf**) overlies the Orwell Limestone and records initiation of deepening of the Appalachian basin and the input of silty and shaly clastic material; it consists of alternating beds of thin-bedded limestone and black calcareous shale. The Glens Falls Limestone was named and described by Ruedemann (1912) and named for Glens Falls, Warren County, N.Y., about 64 km (40 mi) south of Crown Point. The continuation of basin deepening in the Late Ordovician is seen by deposition of calcareous shale of the Stony Point Formation (**Osp**) overlying the Glens Falls Limestone. The Stony Point was named and described by Ruedemann (1912) for occurrences of black shale at Stony Point, Clinton County, N.Y., about 121 km (75 mi) north of Crown Point.

Gamma Radiation Measurements

While mapping the Proterozoic rocks, a portable radiation detector documented background gamma radiation and locally elevated gamma radiation. Locally, elevated readings can be a proxy for identifying rocks with higher thorium (Th) that is associated with elevated rare earth element (REE) concentrations in mineral deposits (Meleik and others, 1978; Force and others, 1982; McCafferty and others, 2014; Shah and others, 2017, 2021). The device (Polimaster PM1703MO-1) has two built-in detectors, a CsI (TI) (cesium iodide thallium activated) scintillation detector and a Geiger-Muller detector. Background readings

averaged about 10 $\mu\text{R/h}$ (microrems per hour), with marble units showing the lowest background of about 5–10 $\mu\text{R/h}$ and the Lyon Mountain Granite Gneiss showing the highest background counts of about 10–15 $\mu\text{R/h}$. Quantitative gamma ray spectrometer data on selected units in the area confirms the semiquantitative gamma radiation readings (Shah and others, 2019, 2021). Continuous data were not collected, and the device was not used at all observed outcrops in the quadrangle because it was not available the entire time of the survey. Readings were recorded in the GIS database when the device detected more than approximately two times background, and those locations are shown on the map with red dots. The greatest number of elevated readings occurred in units **Ybg**, **Ylg**, and **Yp**, and most places with elevated readings contained pegmatite. The average highs and maximum high values for the three map units are provided in [table 3](#).

These data agree with elevated potassium (K) and equivalent thorium (eTh) measurements with a gamma ray spectrometer (Shah and others, 2019, 2021). One location (CP-5010) in an Ediacaran mafic dike (**Zd**) yielded a value of 71 $\mu\text{R/h}$ ([fig. 5](#)). At station CP-4431 (see the GIS database), a thin section of Lyon Mountain Granite Gneiss (**Ylg**) with readings of 25–30 $\mu\text{R/h}$ contains clusters of allanite ([fig. 6](#)). Allanite locally occurs in Adirondack pegmatites and is reported from the Crown Point Spar quarry (Tan, 1966; Lupulescu and others, 2011).

Table 3. Quantitative gamma radiation results on map units from the Crown Point quadrangle.

[Abbreviation: $\mu\text{R/h}$, microrems per hour]

	Map unit		
	Ybg	Ylg	Yp
Maximum ($\mu\text{R/h}$)	78	35	80
Average high ($\mu\text{R/h}$)	37	25	44
Number of measurements	9	13	7



Figure 5. Photographs (A–B) of a joint canyon on the east side of Buck Mountain with remnant fragments of a mafic dike (Zd) exposed on a cliff face of biotite gneiss (Ybg). The Zd dike yielded a gamma radiation value of 71 $\mu\text{R/h}$. Yellow backpack at the base of the cliff is shown for scale. Abbreviation: $\mu\text{R/h}$, microrem per hour. Photographs by Gregory J. Walsh, U.S. Geological Survey.

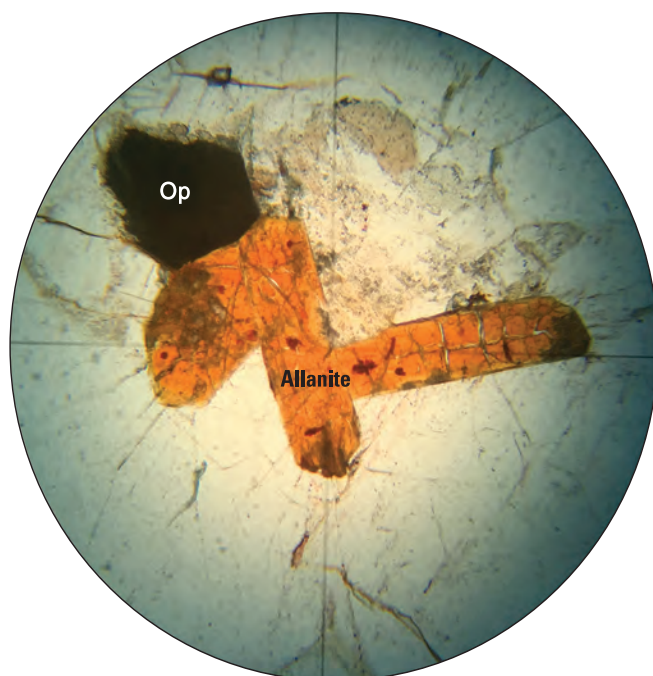


Figure 6. Plane-light photomicrograph showing a cluster of three yellow-orange allanite crystals in a microcline granite sample of the Lyon Mountain Granite Gneiss (Ylg). The rock consists largely of microcline, quartz, and plagioclase with minor amounts of biotite, chlorite, opaques (including magnetite and sulfides), and trace amounts of allanite, zircon, apatite, hematite, and titanite. Gamma radiation readings of this sample were 25–30 $\mu\text{R/h}$, or about 2.5 to 3 times background. Field of view is 1 millimeter. Abbreviations: Op, opaques; $\mu\text{R/h}$, microrems per hour. Photomicrograph image by Gregory J. Walsh, U.S. Geological Survey.

Structural Geology

Ductile Structures

The oldest foliation in the Crown Point quadrangle is a relict gneissosity in the Mesoproterozoic Grenville Complex. The relict gneissosity (S1) is preserved only in the hinge regions of rootless folds (fig. 7). S1 contains layer-parallel leucosome and pegmatite segregations that are foliated and isoclinally folded implying that it developed during a relict high temperature tectonothermal event, which may correspond to an early part of the Shawinigan orogeny (about 1.2–1.14 giga-annum [Ga, billion years before present]; McLelland and others, 2013).

The Mesoproterozoic S1 gneissosity is parallel to a penetrative foliation that is a second-generation (S2) foliation in the Mesoproterozoic rocks. This S2 gneissosity is axial planar to abundant isoclinal and reclined folds in the Grenville Complex. The S2 gneissosity is also the most conspicuous

deformational fabric in the orthogneiss units Ymig, Yggn, and Yhg. Williams and others (2018) report a U-Pb zircon age of $1,177 \pm 10$ Ma from unit Ymig in the adjacent Ticonderoga quadrangle and Regan and others (2019a) report a U-Pb zircon age of $1,185 \pm 11$ Ma from unit Yggn in the adjacent Eagle Lake quadrangle; in both cases the rocks are strongly foliated by S2 implying that the D2 deformation is younger than about 1,177 Ma making it Shawinigan or younger. F2 folds associated with S2 are isoclinal with down plunge fold axes that generally trend subparallel to the L2 mineral lineation and trend east with an average trend of 90° and a plunge of 21° (fig. 7). This trend is considered an average because the D2 fabrics are deformed; poles to the deformed S2 gneissosity show a north-south trending girdle as a result of the younger deformation (fig. 7A), and a slight girdle is present in the L2 data also (fig. 7B).

Both S1 and S2 are deformed by a minimum of two younger ductile fabrics (S3 and S4). The third generation of planar fabrics include upright, open to tight folds (F3) that are only rarely expressed by an axial planar cleavage. The F3 axial surfaces have many different orientations, although they most commonly strike east-northeast and dip steeply (fig. 8). The F3 folds are related to dome and basin formation and are most conspicuous in the core of the Keeney Mountain synform. The large F3 synform extends for over 8 km across the entire southwestern part of the map from Keeney Mountain to Miller Mountain. In the core of the synform, the dominant gneissosity (S2) is subhorizontal, or gently plunging to the southwest. A complimentary upright F3 antiform occurs on the southwest side of Buck Mountain. An overturned F3 synform and upright antiform occur to the northwest passing through Maggie Dudley Road; the overturned F3 synform is consistent with an F3 box-fold geometry seen in some exposures (fig. 8). F3 fold axes have variable plunges with an average trend of 243° , 9° to the southwest (fig. 8). Axial surfaces of F3 folds locally contain thin pegmatite segregations that crosscut S2 (fig. 8). Regional pegmatite ages between about 1,090 to 949 Ma (Lupulescu and others, 2011) and local pegmatite ages in this quadrangle (1,048–1,025 Ma; table 1) indicate that D3 deformation is most likely Ottawaan.

The youngest ductile deformation (D4) in the area occurs as parallel sets of low-amplitude, long-wavelength open folds, boudinage, and shear bands. The dilatant extensional D4 structures are filled with pegmatite (figs. 4D, 9, 10). The boudinage shows variable orientations but includes dominant, conjugate trends that strike north-northeast and west-northwest and dip subvertically (fig. 9). F3 and F4 folds have somewhat different orientations and geometry, but it is not always possible to tell the two fold generations apart in the field. Tighter F3 folds greatly impact the map pattern with east-northeast to east-west folds like the Keeney Mountain synform. Open F4 folds have no impact on the map pattern and show conjugate northwest and more common north-northeast cross-folds, shear bands, and boudinage. Both F3 and F4 folds have pegmatite in their axial surfaces, indicating that they formed at high-grade during similar sillimanite-K-feldspar conditions. Steeply dipping,

northwest and northeast striking shear bands (fig. 3) locally exhibit quartz ribbons and recrystallized feldspar (fig. 10). In one place on the south flank of Bulwagga Mountain, north of Russell Street, an approximately 1-km-long dextral ductile

shear zone was mapped. The S2 foliation northwest and uphill of the shear zone is warped into parallelism with the shear zone near the base of the hillslope. The full extent of the shear zone is not known due to glacial cover.

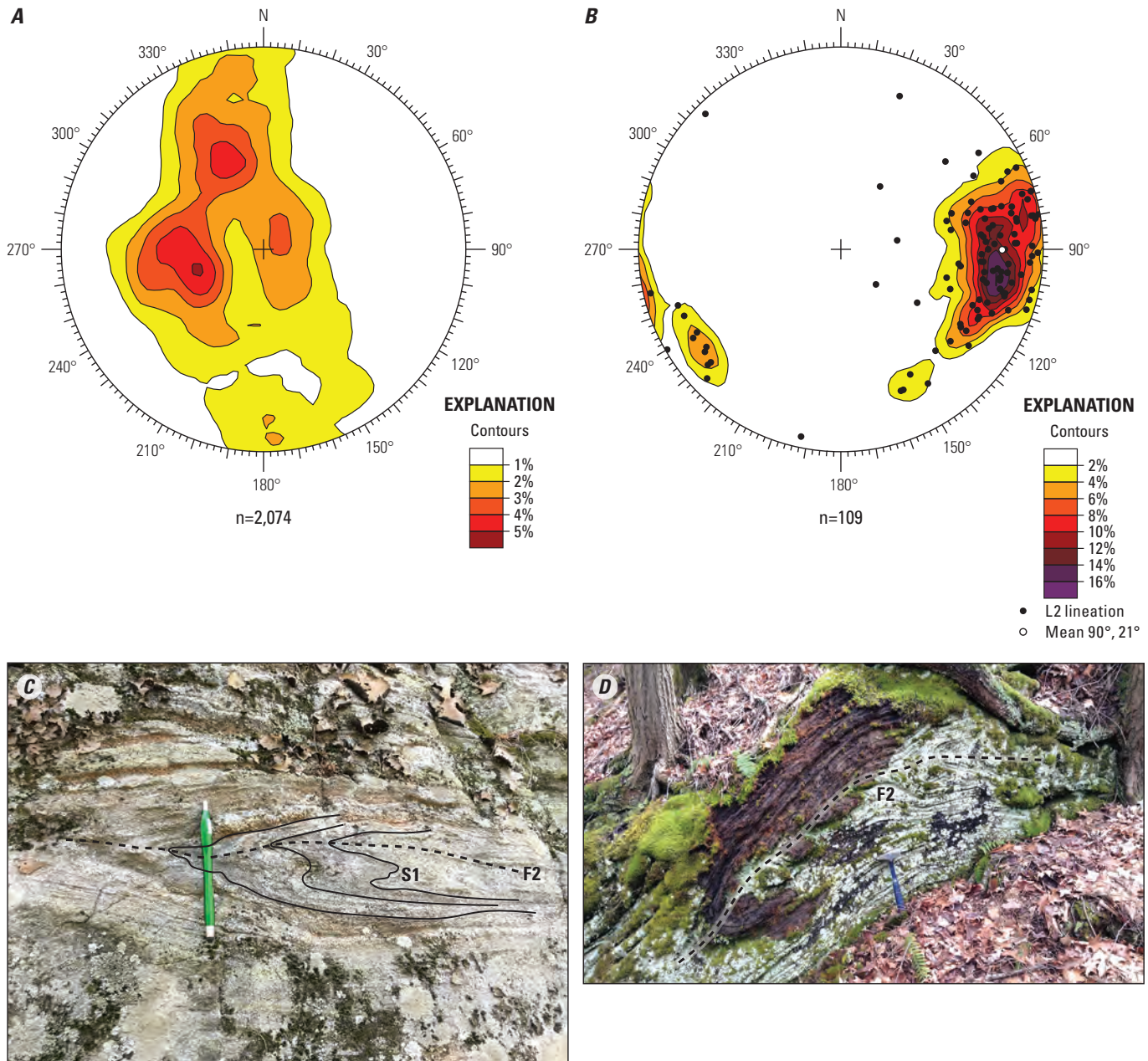


Figure 7. Stereonets (A–B) and photographs (C–D) showing the D2 structures. The stereonets show lower-hemisphere equal area projections of A, contoured poles to S2 gneissosity; and B, contoured L2 mineral lineations and F2 fold axes with a mean trend and plunge of 90°, 21° (see “Mean”). The number of structural measurements in the dataset is indicated by “n” at the bottom of each diagram. Stereonets were plotted using the Structural Data Integrated System Analyser (DAISY 3, version 5.14a) software by Salvini and others (1999) and Salvini (2016). C, photograph of map unit Ybg showing a recumbent, isoclinal, rootless F2 fold and an older S1 migmatitic gneissosity; the axial trace of the F2 fold is shown with a dashed line. D, photograph of marble (Ym) showing a recumbent isoclinal F2 fold warped by a broad upright F3 antiform, with axial surface parallel to the hammer handle, near Buck Mountain Pond in the core of the Keeney Mountain synform. Photographs by Gregory J. Walsh, U.S. Geological Survey.

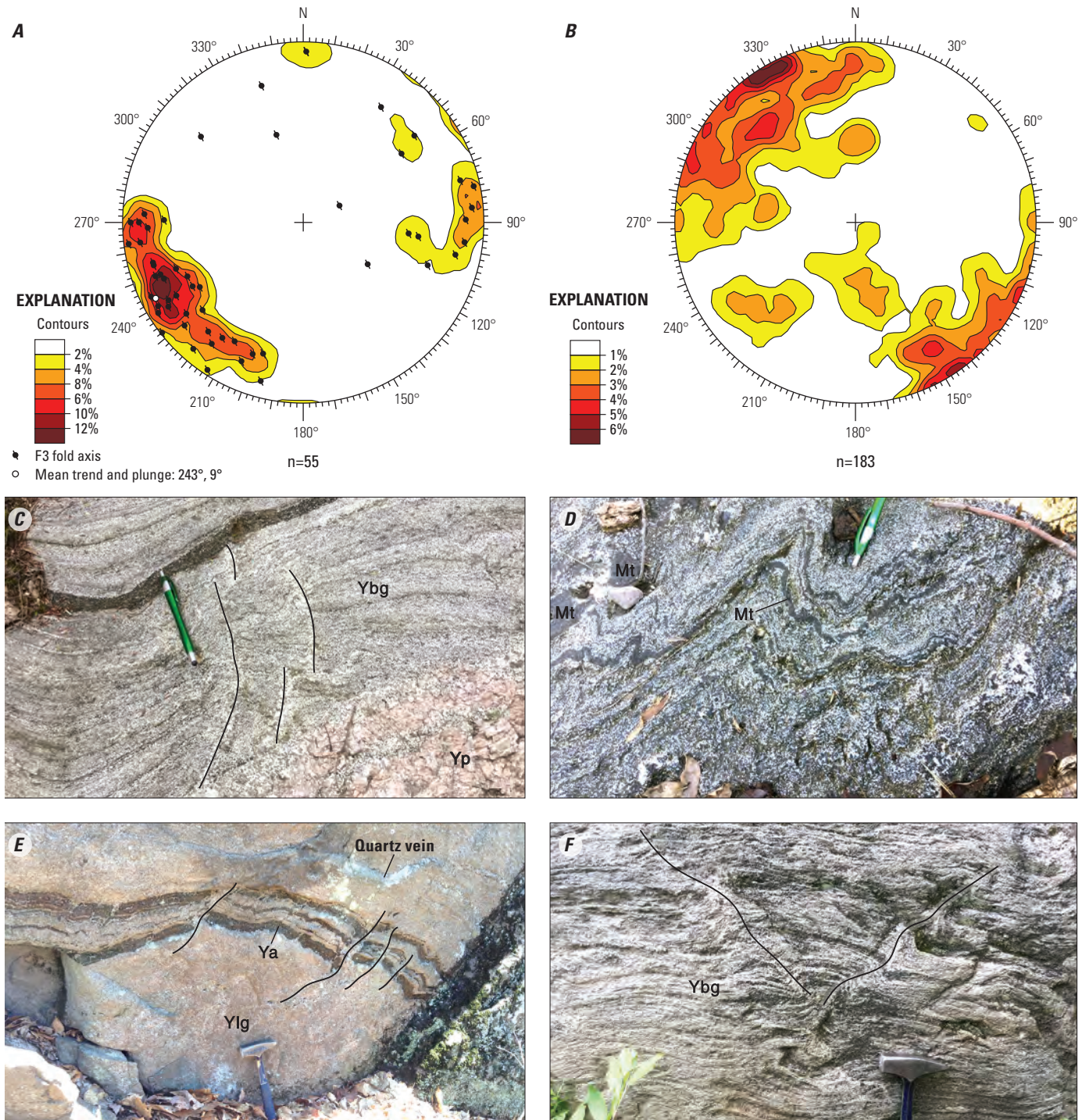


Figure 8. Stereonets (A–B) and photographs (C–F) showing F3 folds and related structures. The stereonet shows lower-hemisphere equal area projections of A, contoured F3 fold axes with a mean trend and plunge of 243°, 9°; and B, contoured poles to F3 axial surfaces showing a principal trend that is steeply dipping with a northeast strike. The number of structural measurements in the dataset is indicated by “n” at the bottom of each diagram. Stereonets were plotted using the Structural Data Integrated System Analyser (DAISY 3, version 5.14a) software by Salvini and others (1999) and Salvini (2016). C, photograph showing pegmatite (Yp) at the lower right and in the axial surfaces of F3 folds (black lines) in unit Ybg. D, photograph showing thin, folded magnetite seams (Mt) and leucocratic layers in unit Ybg. E, photograph showing pegmatite in the axial surfaces of F3 folds (black lines); the S2 foliation is defined by screens of amphibolite (Ya) in unit Ylg; a deformed light-gray quartz vein is visible near the top of the image. F, photograph showing rare box-fold geometry of F3 folds in unit Ybg; the axial traces of the F3 folds are shown by black lines. Pencil in C–D and rock hammer in E–F are shown for scale. Photographs by Gregory J. Walsh, U.S. Geological Survey.

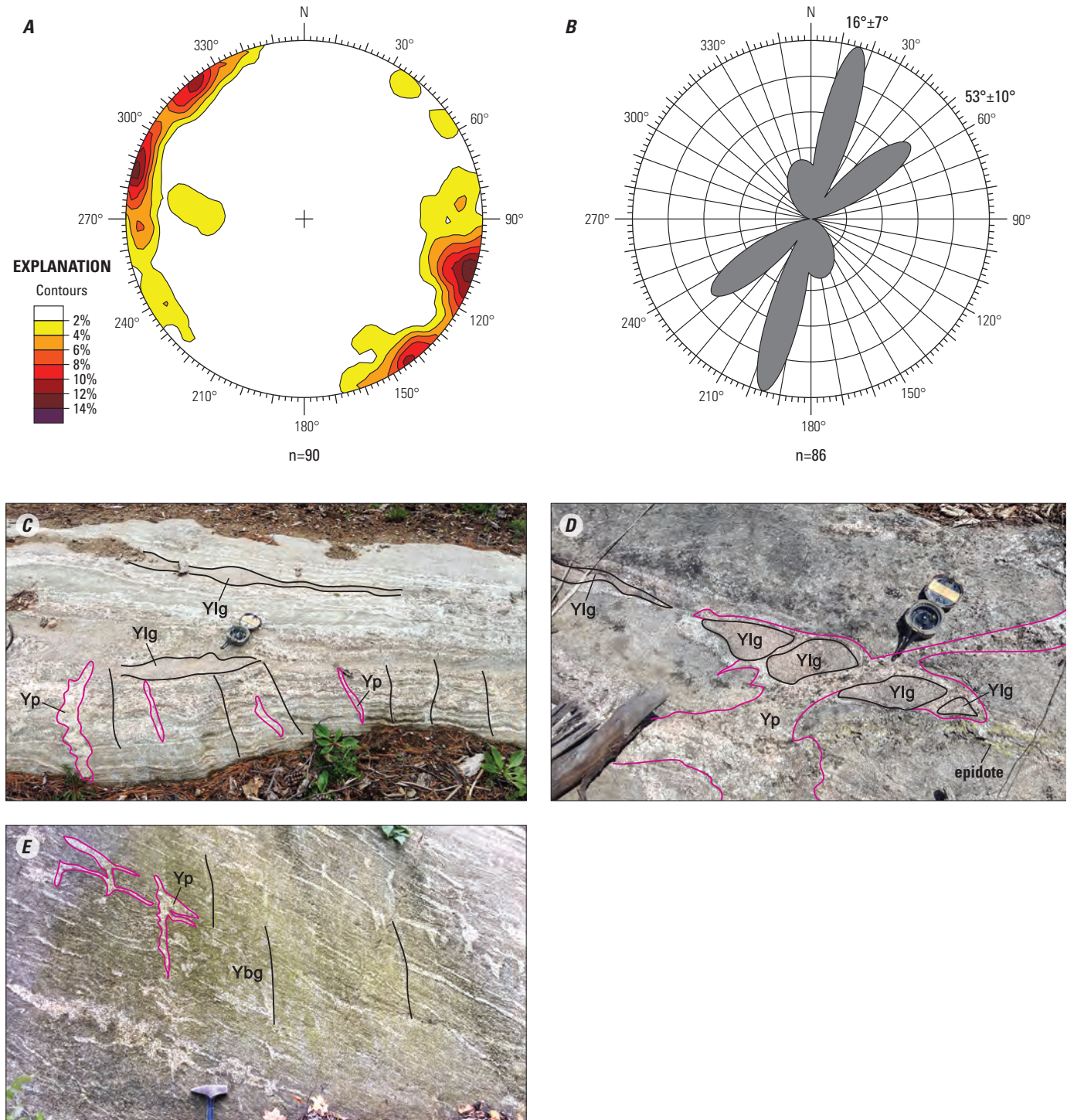


Figure 9. Structural diagrams (A–B) and photographs (C–E) of boudinage or F4 folds. A, stereonet; and B, rose diagram showing the orientations of measured boudinage. The stereonet shows a lower-hemisphere equal area projection of contoured poles to the axial surface of open F4 folds or boudinage. The rose diagram shows a normalized subset of the data for dips $>59^\circ$, and statistical peak trends are shown with 1 standard deviation error (for example, $16^\circ \pm 7^\circ$). The number of structural measurements in each dataset is indicated by “n” at the bottom of each diagram. Stereonet and rose diagram were plotted using the Structural Data Integrated System Analyser (DAISY 3, version 5.14a) software by Salvini and others (1999) and Salvini (2016). C–D, photographs of map unit Ybg showing deformed layers of Ylg, the trace of F4 axial surfaces (subvertical black lines in C), and outlined pegmatite (Yp) in the axial surfaces and dilation zones (pink lines). E, photograph of map unit Ybg showing the trace of F4 axial surfaces (subvertical black lines), and pegmatite (Yp) that occurs in the F4 axial surfaces, in small dilation zones, and as layer-parallel sills (outlined in a few places in pink). Compass in C and D and rock hammer in E are shown for scale. Photographs by Gregory J. Walsh, U.S. Geological Survey.

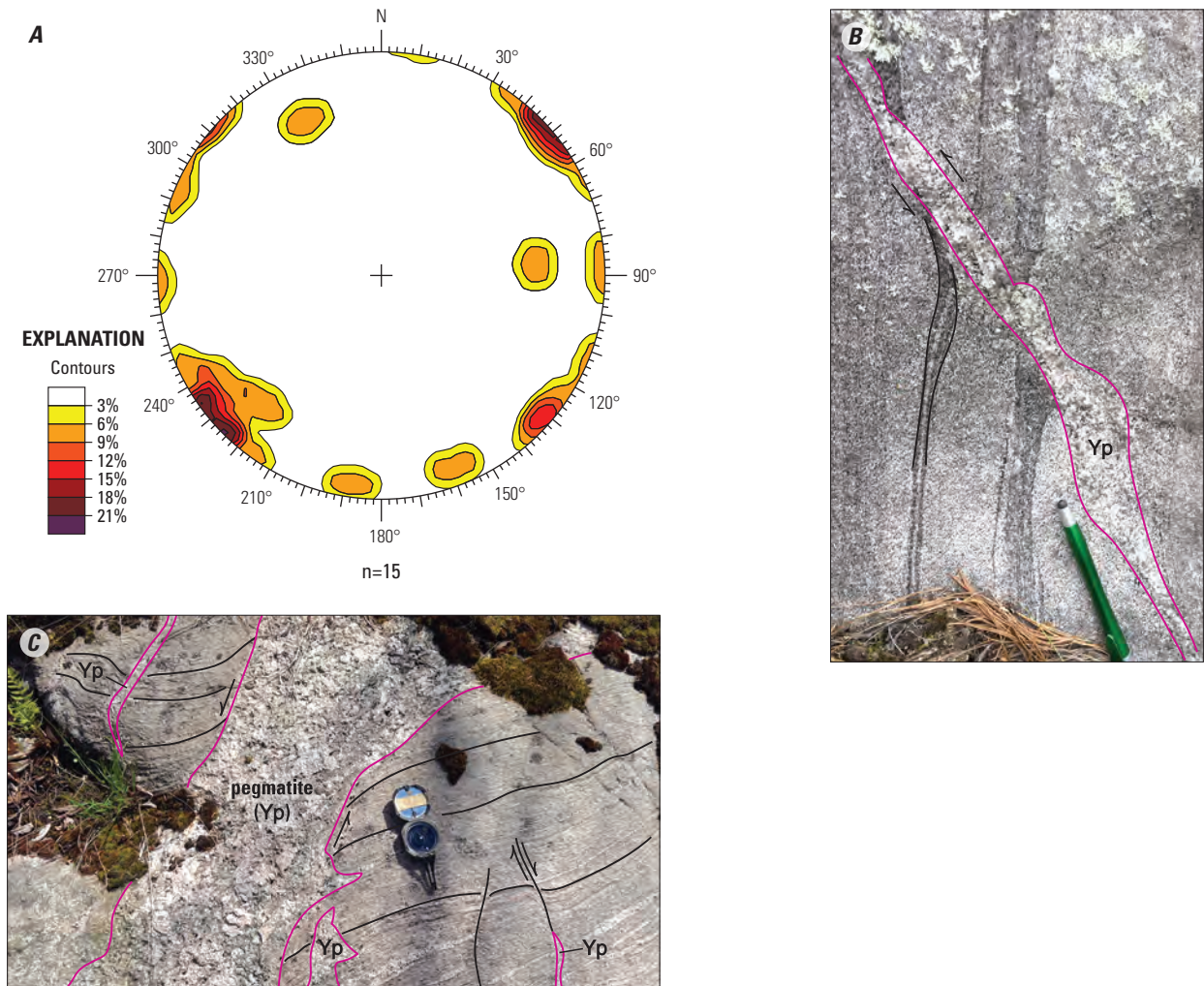


Figure 10. Structural diagram (A) and photographs (B–C) of D4 shear bands or shear zones. A, stereonet showing a lower-hemisphere equal area projection of the orientation of contoured poles to measured shear zones. The number of structural measurements in the dataset is indicated by “n” at the bottom of the diagram. Stereonet was plotted using the Structural Data Integrated System Analyser (DAISY 3, version 5.14a) software by Salvini and others (1999) and Salvini (2016). B–C, photographs showing pegmatite (Yp) filled shear zones (outlined with pink lines); arrows show direction of shearing, and black lines in C show the trace of subvertical shear bands and the trace of the subhorizontal dominant foliation. Pencil in B and compass in C are shown for scale. Photographs by Gregory J. Walsh, U.S. Geological Survey.

Brittle Structures

Northeast striking fractures and lineaments are locally prominent in areas of abundant exposed bedrock. Steeply dipping mafic dikes trend northeast with a mean trend of $60^{\circ} \pm 6^{\circ}$ (fig. 11), matching a regional trend seen on the edge of Laurentia due to the breakup of Rodinia (Burton and Southworth, 2010). The dikes are interpreted as Iapetan rift-related igneous rocks (Coish and Sinton, 1992; Burton and Southworth, 2010). Regionally the dikes display a range of compositions and yield ages analyzed by varying methods including a dike swarm in the Ottawa graben (U-Pb baddeleyite age of $590 \pm 2/-1$ Ma, Kamo and others, 1995); the Callander alkaline complex intrusive body in the Ottawa graben (U-Pb zircon age of 577 ± 1 Ma, Kamo and others, 1995); Rand Hill

diabase dikes in New York (whole-rock K-Ar ages from 588 to 542 Ma, Isachsen and others, 1988); and a trachytic dike at Dannemora, N.Y. (provisional U-Pb zircon age of 643 ± 4 Ma, Chiarenzelli and Regan, 2015).

Veins

Tabular crosscutting veins of quartz were mapped mostly in the Mesoproterozoic rocks. These veins contain mostly quartz and small amounts of epidote, and locally contain calcite. Locally, they occur in the Lyon Mountain Granite Gneiss as tabular sills, parallel to the layering, or crosscutting veins, and these veins are likely much older and related to pegmatite and iron-ore seam emplacement. Epidote veins locally show

marginal bleached zones due to hydrothermal alteration (fig. 12), and saussurite alteration of plagioclase is visible in thin section. The veins show a variety of orientations dominated by steeply dipping north-south and east-west trends (fig. 12).

Joints

Summary stereonet and rose diagrams show the structural orientations of over 2,000 measured outcrop-scale joints across the map area (fig. 13). Most measured joints are steeply dipping and north-northwest striking, and a small subset of joints shows gentle dips (fig. 13A). The strike and dip of outcrop-scale joints are not plotted on the bedrock geologic map but are included in the GIS database. The principal trend of steeply dipping joints is $347^{\circ} \pm 18^{\circ}$ with a subordinate trend at $282^{\circ} \pm 20^{\circ}$, and a minor northeast trend at $48^{\circ} \pm 17^{\circ}$ (fig. 13B). Many of the joints are visible as lineaments on the percent-slope map in areas with significant exposed bedrock, such as Buck Mountain and Miller Mountain (see bedrock geologic map).

Faults

Brittle faults are characterized by breccia, quartz-calcite veins, and rarely observed pseudotachylyte. Asbestiform serpentine minerals occur on fault surfaces in marble due to the retrogression of olivine (fig. 14). The principal trends of measured brittle faults are steeply dipping east-west, northeast, and north-south striking (fig. 15). The strike and dip of outcrop-scale faults are not plotted on the bedrock geologic map but are included in the GIS database. The same trends are seen in the mapped brittle faults. Faults with generally short trace lengths of approximately 1 km were mapped by the surface expression of lineaments on the lidar percent-slope map and by observed minor offset of geologic units. It is likely that many more linear features seen in the lidar (see simplified surficial geologic map), but not mapped in detail, may represent brittle faults with very minor offsets or zones of abundant joints with minimal to no offset. Large trace-length faults, especially the concealed faults beneath the glacial cover and Lake Champlain are inferred from offset Paleozoic stratigraphy. Apatite fission track and thermochronology data indicate that tectonic unroofing of the Adirondacks took place from the Middle Jurassic through the Late Cretaceous (about 170 to 70 Ma; Roden-Tice and others, 2000; Roden-Tice and Tice, 2005). Fission track and U-Pb calcite ages indicate that the passive margin experienced renewed tectonic activity at about 105, 85 to 65, and 20 Ma and these may correlate with timing of renewed brittle faulting (Amidon and others, 2016; Robbins and Amidon, 2018). Recent work shows that calcite veins in the Champlain Valley yield U-Pb ages that document fracturing from about 115 to 75 Ma followed by a period in the Miocene to Pliocene (Amidon and others, 2022), suggesting that brittle tectonics were active from the Late Jurassic to the Neogene. The entire Adirondack dome is experiencing anomalously high uplift (Isachsen, 1981).

Champlain Valley Faults

Normal faults mapped in the Champlain Valley are likely related to Mesozoic extension (Selleck, 1980) as they show consistent stress orientations with Mesozoic faults in other parts of New England. Welby (1961) and Selleck (1980) also stated that these faults may be as old as Ordovician and were reactivated at various times. Although these faults are largely covered by glacial deposits and Lake Champlain, major changes of lithostratigraphic units and gentle dips of bedding over large distances require their existence. Also, fault breccia exists in some outcrops in the Crown Point quadrangle and demonstrates the high-angle nature of these structures. The major change from Upper Cambrian and Lower Ordovician rocks on the west shore of the lake to Upper Ordovician rocks on the eastern shore is evidence of a normal fault within Lake Champlain that was recognized by Welby (1961). For example, on the western shore, rocks of the Upper Cambrian to Lower Ordovician Whitehall Formation are exposed just south of the town of Crown Point at Rock Way (name not on map) near Sheepshead Island and dip to the northeast, whereas on the eastern shore, rocks of the Upper Ordovician Glens Falls Limestone are exposed just north of Leonard Bay, Vt., and dip to the northwest. This change from older to younger rocks from west to east suggests a north-south trending, down-to-east normal fault for the Lake Champlain fault and is also recognized in the southeastern part of the quadrangle where it is offset to the east.

The sparseness of outcrops due to glacial cover in the Champlain Valley makes it difficult to locate faults, a point also noted by Welby (1961). Faults along the railway-cut south of Hickock Point, N.Y., show normal and strike slip motion. The faults are locally characterized by breccia and locally offset beds. The Schoolhouse Bay fault is an oblique, right-lateral strike-slip fault that was recognized along the railroad cuts. It extends to the southwest where it marks the boundary between the uplifted basement rocks to the west and the down-dropped and laterally displaced rocks of the Champlain Valley to the east; however, the fault is largely inferred because it is not exposed to the south. The Vineyard Road fault is not exposed but is inferred from a slight normal offset of the basement rocks in the valley where Vineyard Road passes near the abandoned Vineyard mine. Further north, the inferred Vineyard Road fault shows greater normal displacement along the east side of Bulwagga Mountain suggesting a wrench-fault geometry. The Leonard Bay fault in Vermont is also inferred. The faults are consistent with observations made by Rodgers (1937) who noted four sets of high-angle faults in the Lake Champlain region: (1) longitudinal faults with strikes ranging N-S to N 40° E; (2) transverse faults with strikes ranging from N 85° W to N 75° E; (3) northwest minor faults with average strike of N 45° W; and (4) northeast minor faults with average strike of N 45° E. The high-angle normal faults in the Crown Point quadrangle mostly fall into his longitudinal and transverse fault categories, and generally match the observed directions seen in our mapped and outcrop-scale faults (fig. 15).

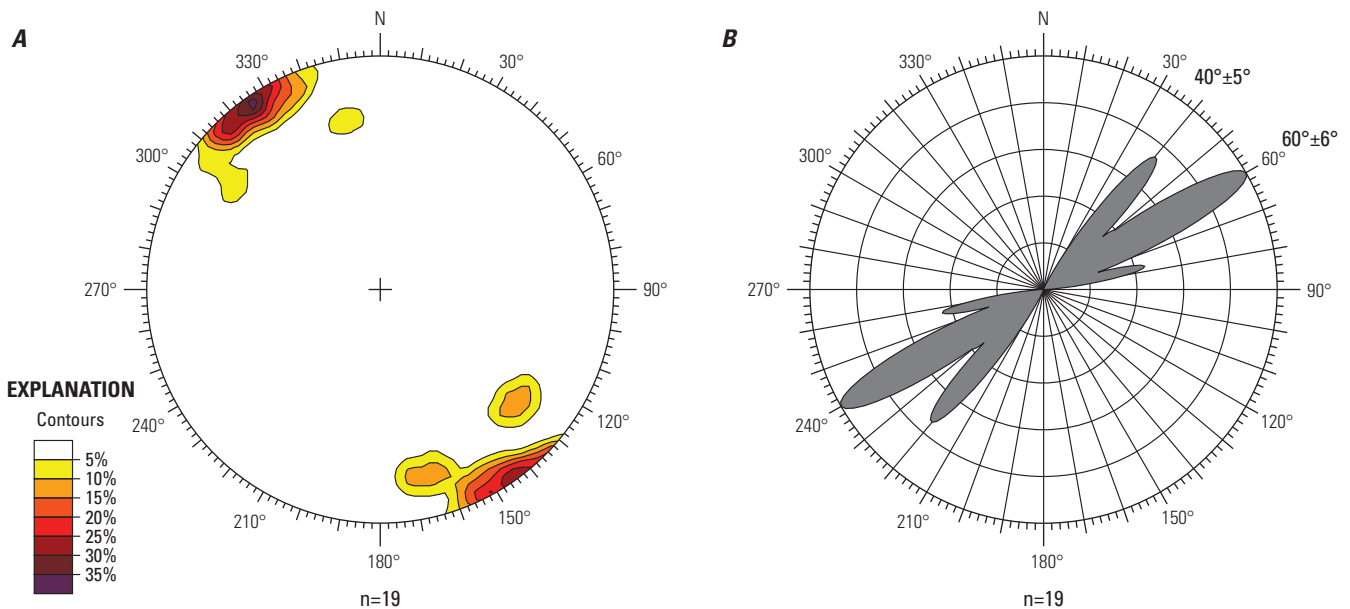


Figure 11. Structural diagrams (A–B) showing the orientation of Ediacaran mafic dikes (Zd). A, stereonet; and B, rose diagram showing the orientation of measured mafic dikes. The stereonet shows a lower-hemisphere equal area projection of contoured poles to dikes. The rose diagram shows strike data for all dikes since all dips are >59°; statistical peak trends are shown with 1 standard deviation error (for example, 60°±6°). The number of structural measurements in each dataset is indicated by “n” at the bottom of each diagram. Stereonet and rose diagram were plotted using the Structural Data Integrated System Analyser (DAISY 3, version 5.14a) software by Salvini and others (1999) and Salvini (2016).

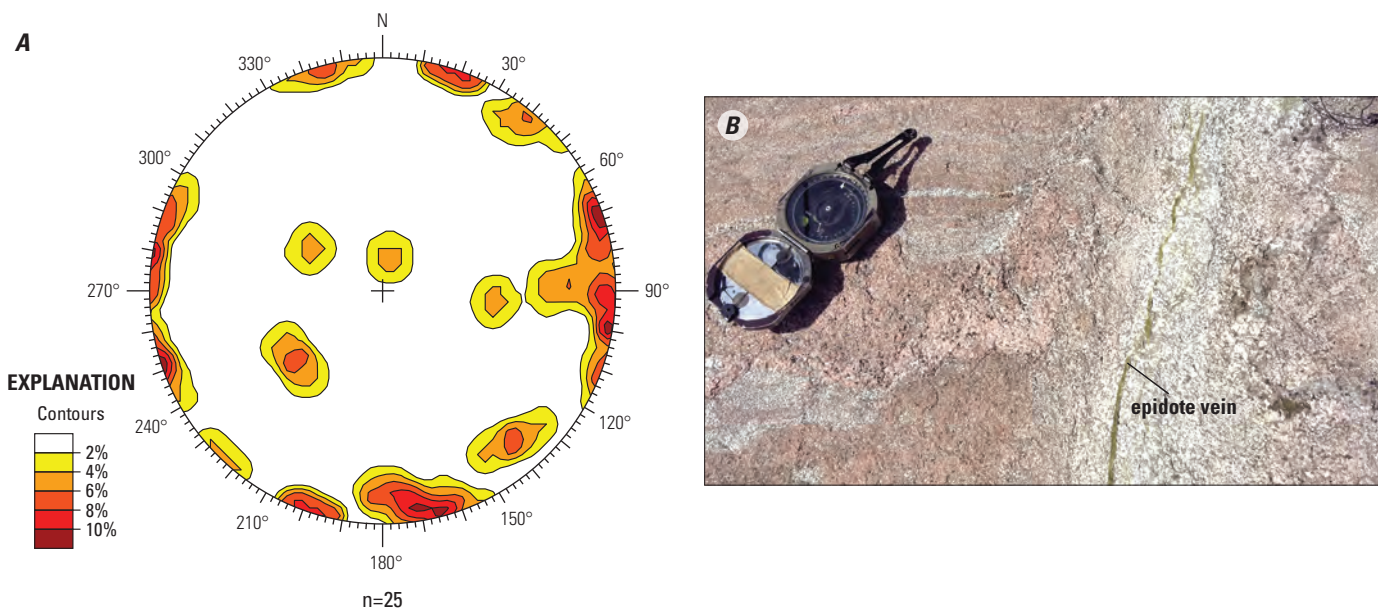


Figure 12. Stereonet (A) and photograph (B) showing veins. A, stereonet showing a lower-hemisphere equal area projection of contoured poles to veins. The number of structural measurements in the dataset is indicated by “n” at the bottom of the diagram. Stereonet was plotted using the Structural Data Integrated System Analyser (DAISY 3, version 5.14a) software by Salvini and others (1999) and Salvini (2016). B, photograph showing an epidote vein in unit Ylg. Compass is shown for scale. Photograph by Gregory J. Walsh, U.S. Geological Survey.

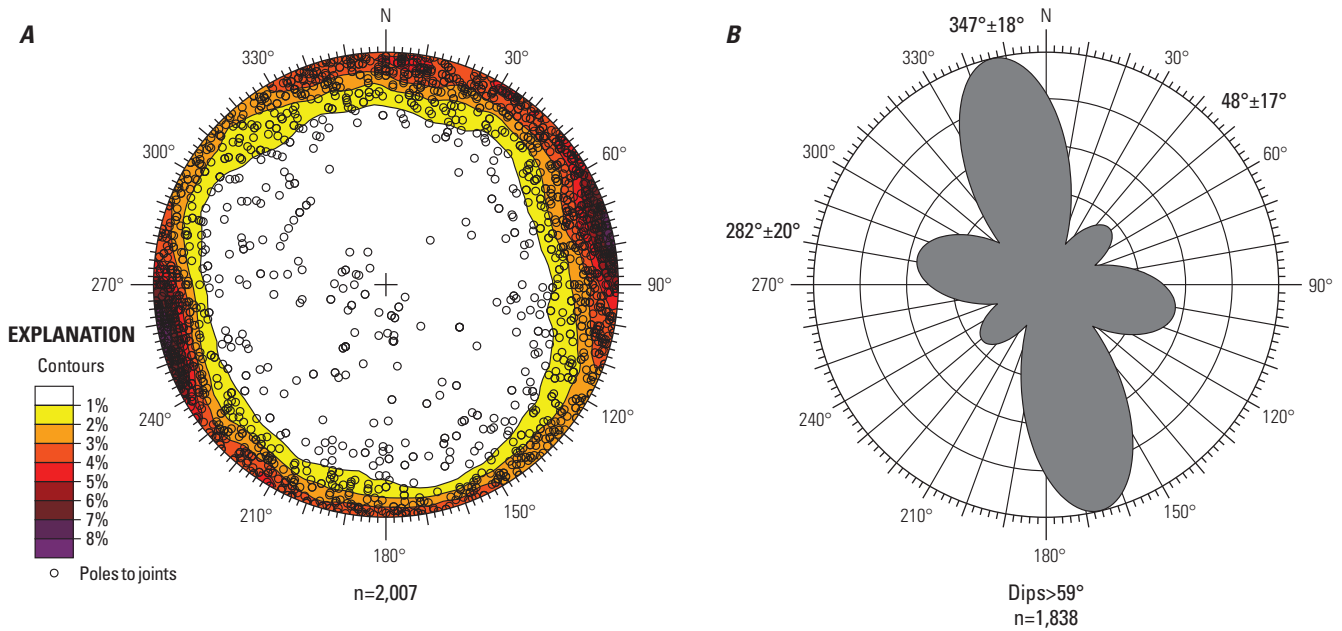


Figure 13. Stereonet (A) and rose diagram (B) showing the orientations of measured joints. A, stereonet showing a lower-hemisphere equal area projection of poles and contoured poles to joints. B, rose diagram showing a normalized subset of the data for dips >59°, and statistical peak trends are shown with 1 standard deviation error (for example, 347°±18°). The number of structural measurements in each dataset is indicated by “n” at the bottom of each diagram. Stereonet and rose diagram were plotted using the Structural Data Integrated System Analyser (DAISY 3, version 5.14a) software by Salvini and others (1999) and Salvini (2016).

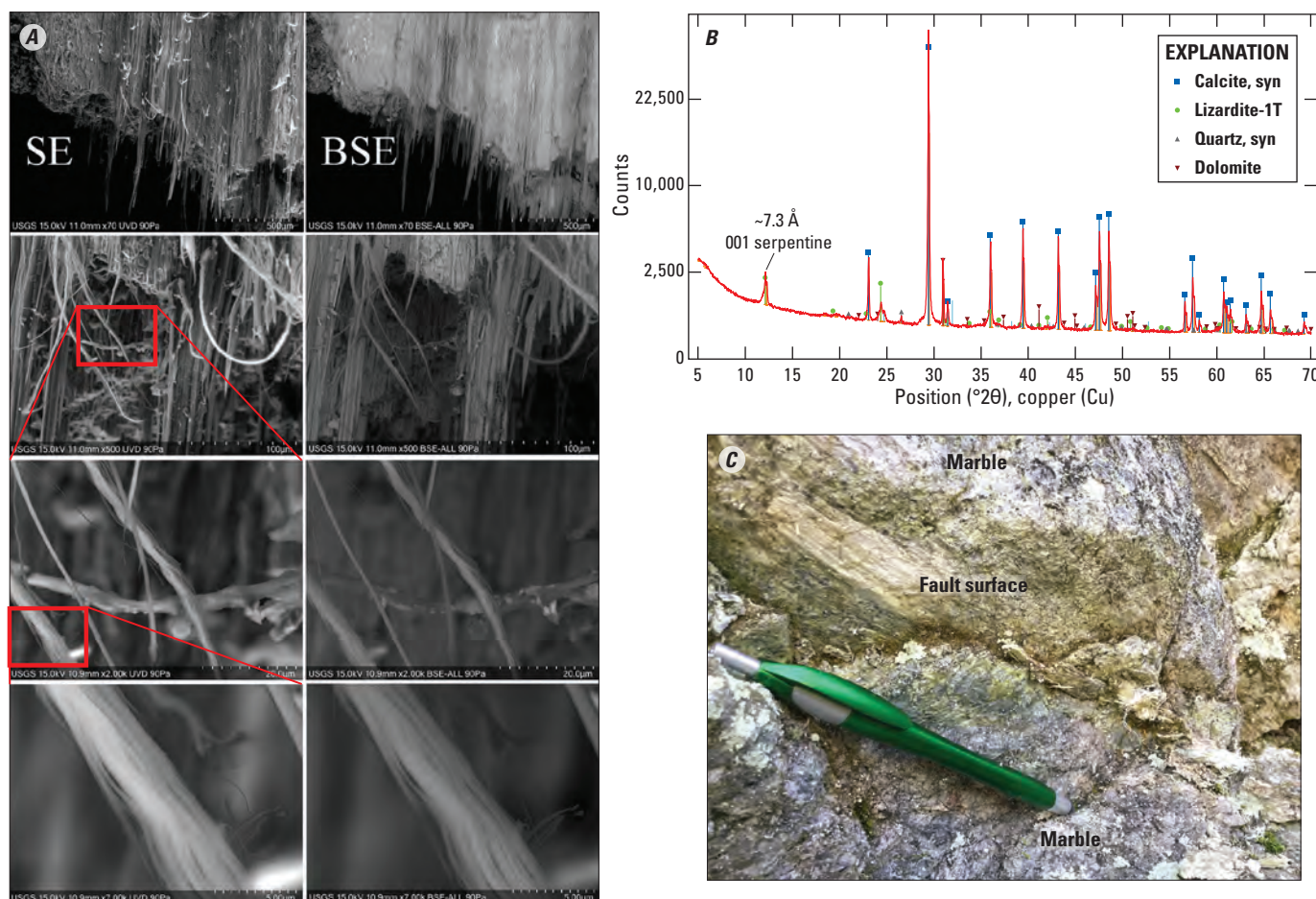


Figure 14. SEM images (A), powder X-ray diffraction (XRD) spectrum (B), and photograph (C) of a fracture- and fault-filling serpentine mineral from a marble sample collected from the Buck Mountain Pond graphite mine on Mine Hill (sample CP-5048A). A, scanning electron microscope (SEM) images at progressively higher magnifications from top to bottom; secondary electron images (SE, left column, top to bottom), and backscattered electron images (BSE, right column, top to bottom). The SEM images show that individual fibers are comprised of bundles of sub-500-nanometer fibers. Aspect ratios of individual fibers are >100 . The SEM images were taken on a variable pressure Hitachi SU-5000 field emission electron microscope (FE-SEM) at a chamber pressure of 90 pascals (Pa) and an accelerating voltage of 15 kilovolts (kV). B, powder X-ray diffraction (XRD) spectrum showing a strong reflection at approximately 7.3 angstroms (Å), consistent with a serpentine mineral exhibiting cleavage on the 001 plane. C, photograph of marble (Ym) rock sample showing fibrous mineral habit on a tan-weathering fault surface suggesting it is a chrysotile polymorph. Using the definition of Campbell and others (1977), the chrysotile is defined as "asbestiform." Coordinates of the sample in decimal degrees are 43.901282, -73.470685 (WGS 84 datum). Pencil in C is shown for scale. Abbreviations: BSE-ALL, backscattered electron image with all segments of the detector active; k, thousand; mm, millimeters; syn, synthetic; 1T, 1T trigonal polytype of lizardite; 2θ , 2-theta position in degrees (sample run with Cu X-ray source); μm , micrometers; USGS, U.S. Geological Survey; UVD, ultra variable-pressure detector. Photograph by Gregory J. Walsh, U.S. Geological Survey.

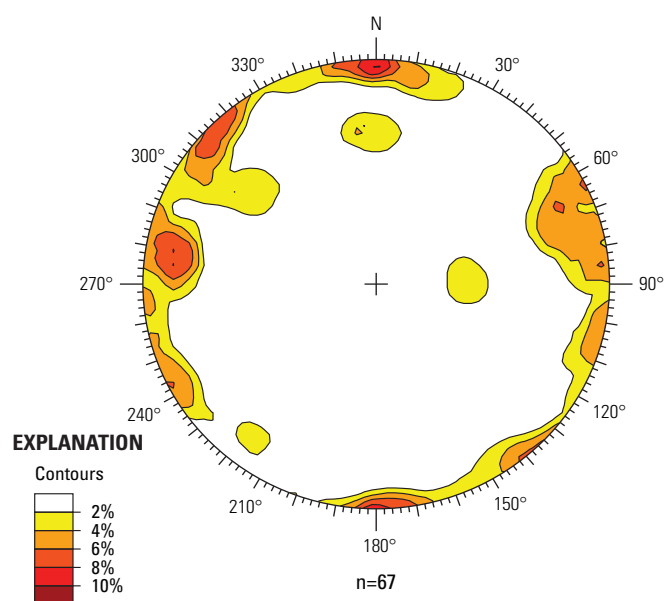


Figure 15. Stereonet showing the orientations of measured brittle faults. The stereonet shows a lower-hemisphere equal area projection of contoured poles to faults. The number of structural measurements in the dataset is indicated by “n” at the bottom of the diagram. Stereonet was plotted using the Structural Data Integrated System Analyser (DAISY 3, version 5.14a) software by Salvini and others (1999) and Salvini (2016).

Tectonics and Metamorphism

Tectonic and metamorphic events in the Adirondack Highlands of the Grenville Province are associated with three major events called the Elzevirian (about 1.25–1.22 Ga), Shawinigan (about 1.2–1.14 Ga), and Ottawa (about 1.09–1.03 Ga) orogenies (McLelland and others, 2013). The last, or fourth, event has been called the Rigolet phase (about 1.01–0.98 Ga) of the Grenvillian orogeny (Rivers, 2012; McLelland and others, 2013). Plutonic calc-alkaline arc-related rocks related to the Elzevirian orogeny occur in the southern Adirondacks and the Mount Holly Complex in the Green Mountains (Ratliff and others, 1991, 2011; Aleinikoff and others, 2011; McLelland and others, 2013) but they do not occur in the Crown Point quadrangle. The multiple high-grade metamorphic events and polyphase ductile deformation have challenged researchers working on the Grenville Province for well over a century (Rivers, 2015). In the Adirondack Highlands, Walton and de Waard (1963, p. 98) noted that “metamorphism in the hornblende or pyroxene granulite facies strongly modified all preexisting rock relationships and totally recrystallized most of the rocks.” They suggested that much of the geologic history of the Adirondack Mountains was, “hidden behind a screen of diastrophism.” The Adirondack Highlands experienced peak metamorphism at granulite facies followed by upper amphibolite facies conditions. The timing of peak granulite facies conditions places it during the

Ottawan orogeny from about 1,080 to 1,070 Ma (McLelland and others, 2001). Earlier high-grade metamorphism occurred during the Shawinigan orogeny. Recent work by Williams and others (2019) shows that the rocks in the eastern part of the Adirondacks Highlands experienced at least two periods of melting, related to biotite dehydration, with the first major event at about 1,160 to 1,150 Ma (Shawinigan orogeny) and a second high-temperature event with less melting at about 1,050 Ma (Ottawan orogeny). This polyphase metamorphism is also documented in Adirondack anorthosite and rocks along the margin of the Marcy massif (Peck and others, 2018; Regan and others, 2019a). Lupulescu and others (2011) dated two layer-parallel pegmatites in Crown Point and reported ages of $1,048 \pm 14$ Ma and $1,025 \pm 3$ Ma, which they interpret as representing granitic melts produced at the end of the Ottawa orogeny. Metamorphism and fluid alteration took place from the Ottawa orogeny to the Rigolet phase of the Grenville orogeny, and the ages of mineral growth are documented by zircon, garnet, monazite, and titanite dated from about 1,050 to 980 Ma (McLelland and others, 2001, 2002; Valley and others, 2009, 2010, 2011; Aleinikoff and Walsh, 2015, 2016, 2019; Aleinikoff and others, 2018; Peck and others, 2018; Regan and others, 2019a, b).

Mineral assemblages in the paragneiss units are consistent with lower granulite facies to upper amphibolite facies conditions, sometimes called the clinopyroxene-almandite subfacies of the granulite facies (de Waard, 1965). Representative assemblages include the following:

- Pelitic and psammitic rocks: garnet-sillimanite-biotite-plagioclase-K-feldspar-quartz-graphite.
- Marble: diopside-forsterite-scapolite-calcite-dolomite-wollastonite-graphite.
- Mafic rocks: garnet-hornblende-clinopyroxene-plagioclase±orthopyroxene.
- Granitoid gneisses: garnet-hornblende-biotite-clinopyroxene-plagioclase-K-feldspar-quartz.

Sillimanite occurs aligned with both L2 mineral lineations and F3 fold axes indicating second sillimanite conditions during both these events. The late pegmatites do not contain muscovite; they crystallized mainly above muscovite stability and can be classified as abyssal pegmatites related to adiabatic melting during uplift and unroofing (Cerny and Ercit, 2005). The granitic gneisses commonly show blebs or planar to lenticular microperthite lamellae textures, which are common in granulite facies rocks where temperatures peak above the alkali feldspar solvus and two feldspars exsolve during slow cooling (Hyndman, 1972). Locally “two stage” microperthite texture with blebs and lamellae occur, and these are characteristic of granulite facies rocks (Parsons and others, 2005). Common trace minerals include apatite, titanite, monazite, magnetite, pyrite, graphite, and zircon. Greenschist facies minerals mark retrograde assemblages. Muscovite is absent except in rocks that show retrograde assemblages, which are

most common as muscovite replacing sillimanite or alkali feldspar, or fine-grained sericite in saussurite. This indicates that all the Mesoproterozoic rocks were metamorphosed above the muscovite stability curve or the activity of water was low in these dry rocks (Storm and Spear, 2005; Williams and others, 2019). Chlorite is a common retrograde mineral after biotite, garnet, and amphibole but retrograde assemblages are not widespread. Epidote is a common retrograde mineral and is locally found in veins; plagioclase is commonly saussuritized to varying degrees where observed in thin section. Actinolite is locally observed, and serpentine minerals are retrograde products of olivine in marbles (fig. 14). Tremolite replaced diopside in green calc-silicate granofels (Ycs).

Economic Geology

Several small abandoned hard rock mines and quarries occur in the area (fig. 16 on map sheet). Small prospects also occur and are included in the GIS database but are not shown on the map or in figure 16. In this study, we attempted to accurately locate and field check all the hard rock mines (fig. 16, table 4 on map sheet). The current USGS Mineral Resources Data System (MRDS) database lists many of the locations incorrectly (Mineral Resources Data System, 2005). The USGS MRDS also lists multiple sand and gravel pits in the area. A few pits were active at the time of mapping, but the Quaternary geology remains to be mapped and we did not field check or map the sand and gravel pits. The largest active pits are located in the Street Road delta and they are described by De Simone and others (2008). The approximate location of the delta is labelled on the bedrock geologic map, and the delta and pits are visible on the lidar (see simplified surficial geologic map). The Pleistocene delta was built from streams draining the Adirondacks into glacial Lake Vermont, and kettles now exposed on the top of the delta demonstrate that it buried blocks of ice (De Simone and others, 2008). The kettles are visible on the topographic base map (see bedrock geologic map) and the lidar (see simplified surficial geologic map).

Iron Mines

Historic iron mines occur in magnetite deposits categorized as iron oxide-apatite (IOA) end-member deposits of the low-Ti Fe-oxide-copper-gold (IOCG) system (Foose and McLelland, 1995; McLelland and others, 2002; Valley and others, 2009; Taylor and others, 2019). The magnetite deposits were mined for iron in the 1800s and supplied the furnaces at Crown Point on Lake Champlain (figs. 16 and 17 on map sheet) (Newland and Kemp, 1908). Iron mining in the Adirondacks significantly impacted the history of the United States including the Revolutionary War, the birth of the U.S. Navy, the development of the railroads and canals, the Industrial Revolution, and the first and second World Wars (Barker, 1942; Farrell, 1996). The main period of iron mining

in the Adirondacks occurred from 1860 to 1880 (Newland and Kemp, 1908). The history of the iron industry in Crown Point is described by Witherbee (1906), Newland and Kemp (1908), Newland (1921), and Barker (1942). According to Barker (1942), in 1872, the Crown Point Iron Company built a new narrow-gauge railroad and two blast furnaces; the latter located on the western shore of Lake Champlain, just east of downtown. Barker (1942) states that the lakeshore area near the current boat launch was the site of a wharf, railyard, two blast furnaces, and related infrastructure to support the iron industry; the blast furnaces are shown in figure 16. The lakeshore area is currently underlain by mine waste and related fill material. The railroad connected the lakeshore to the western upland operations at Ironville and the mines at Hammondville in the adjacent Eagle Lake quadrangle. Peak operation and profitability for the Company took place between 1876 and 1882. The two furnaces on Lake Champlain processed considerable ore, not just from local sources, but from as far away as the Chateaugay mines in Clinton County. In 1900, the railroad was removed, and in 1905 the furnaces were dismantled in response to changing economic conditions and the citing of new iron deposits west of Lake Superior. The deposits and historic workings are now greatly modified by time, and in many places, overgrown by dense forest. Locating and mapping of the abandoned mines, waste rock piles, and the trace of the railroad bed were made possible by recently acquired lidar topographic data.

The iron deposits are hosted in the migmatitic biotite gneiss member of the Grenville Complex (Ybg) and leucogranite of the Lyon Mountain Granite Gneiss (Ylg). The deposits occur as veins (0.5 centimeters [cm] to several meters thick) that are in most places semiconcordant to the principal layering in the host rocks (fig. 18). Crosscutting magnetite veins and magnetite-rich pegmatite also occur. The magnetite deposits within paragneiss, like at the Vineyard mine, locally contain appreciable amounts of pyrite, which made them less suitable for the iron furnace (Newland and Kemp, 1908).

The Vineyard and Butler mines, west of Route 7 (Vineyard Road), were thought to be part of the same deposit by Newland and Kemp (1908). Our mapping shows that there are several pits in the Vineyard mine area that occur in migmatitic biotite gneiss (Ybg), syenitic gneiss, crosscutting tabular magnetite veins, and magnetite-rich pegmatite. The small Butler mine is in a thin ore seam within pink leucogranite (Ylg). Newland and Kemp (1908, p. 41) reported that the ore at the Vineyard and Butler mines occurred in "laminated black hornblende gneiss," and we noted small amounts of amphibolite in the area, but it does not appear to be the main host rock. The Breed and Hammond mines (Newland and Kemp, 1908) occur on the north side of Breeds Hill. Both are located within leucogranite (Ylg) that contains screens and layers of biotite paragneiss (Ybg). Three pits occur within leucogranite (Ylg) on the southeast slope of Miller Mountain at the Kent mine. Two unnamed mines were located in this study, but information matching their locations could not be confirmed in the literature; one occurs in leucogranite

(Ylg) near the contact with biotite paragneiss (Ybg) 1.2 km south-southwest of Crown Point Center, and the other occurs in leucogranite (Ylg) interlayered with biotite paragneiss (Ybg) and marble and calc-silicate rock (Ym) about 3.5 km northwest of Crown Point Center.

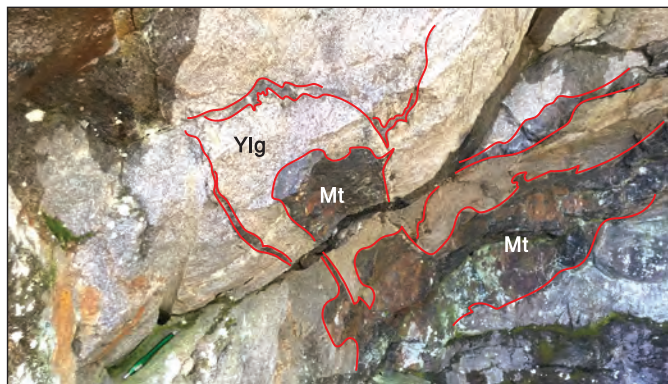


Figure 18. Photograph of magnetite (Mt) ore seams at the Kent mine in host leucogranite gneiss (Ylg). Here, the dark ore seams (outlined in red) occur in veins (1- to 30-cm thick) that are subparallel to the folded foliation (S2) and crosscut the foliation. The veins are deformed by upright F3 folds. Small mushroom-shaped diapirs of ore (right center) locally protrude into the host leucogranite gneiss (Ylg). The ore locally contains pyrite, which results in a rusty stain. Green pen (lower left) is shown for scale. Photograph by Gregory J. Walsh, U.S. Geological Survey.

Graphite

Graphite deposits were mined in the Adirondacks beginning in the mid-1800s (Alling, 1918). Several small graphite mines and prospects were in operation in the area, but all were abandoned by the summer of 1917 (Alling, 1918). The USGS MRDS lists sites named Buck Mountain Pond, Mason, Towne, Kirby, and Mason (fig. 16, table 4). The observed deposits are flake graphite associated with marble; graphite flakes up to 2 cm across are common in the deposits. The “Kirby graphite mine” (Whitlock, 1903) was not located in this study. Whitlock (1903) reports the Kirby location as 3 miles northwest of Ticonderoga, but Alling (1918) did not use the name “Kirby” in his report and this site remains unconfirmed.

Feldspar

Economic feldspar came from pegmatite bodies. The USGS MRDS lists sites named Rose Rock and Crown Point Spar (fig. 16, table 4). The Rose Rock quarry was also named the Buck Mountain quarry and the rose color of the feldspar is due to hematite (Tan, 1966). The Crown Point Spar Company quarry was once the largest and one of the most important producers in New York (Newland, 1921; Tan, 1966). Other

names for the quarry include the Breeds Hill pegmatite, and Spar Mine (Tan, 1966). Tan (1966, p. 53) shows a detailed map of the Crown Point Spar Company quarry on the north slope of Miller Mountain. According to Tan (1966) the mill was built in 1906, and the material from the quarry was used to produce enamel, ornamental stone, and chicken grit. The overhead tramway operated from the quarry to the base of Breed Hill near Lake Champlain and the Delaware and Hudson railroad (Newland, 1921; Tan, 1966). Vestiges of the concrete footings for the tramway, and at least one ore cart, are located on the forested hillside, and the concrete silos at the old mill site are still visible along Route 22/9N, west of Spar Mill Bay.

Paleozoic Rocks

The Paleozoic sedimentary rocks (fig. 19 on map sheet) host abandoned aggregate quarries (fig. 16 and bedrock geologic map), which occur in the Crown Point Limestone (Ocp) and Providence Island Dolomite (Opi) along Grant Brook near the abandoned Champlain Speedway, and in the Cutting Dolomite (Ocu) 0.6 km south-southeast of the junction of Route 22/9N and 185 (labelled “Route S” on the base map).

Phosphate

A phosphate mineral called eupyrchroite, an amorphous calcium phosphate closely resembling apatite, was produced for fertilizer on a small scale around 1850 at a location just south of Crown Point near the shore of Lake Champlain (Newland, 1921) near the Crown Point landing (Emmons, 1838). Newland (1921, p. 27) writes, “It is mentioned in the reports of the First Geological Survey [Emmons, 1838], and mining operations were carried on for a time about the year 1850 with the production of 100 tons or so of the material for use as fertilizer.” Jackson (1851) reports that analyzed samples came from Mr. C.F. Hammond of Crown Point, and 100 tons were removed from a mine for agricultural phosphates. The mineral is hosted in gneiss (Emmons, 1838). The location of the source mine was not confirmed in this study; it may be located along the Sugar Hill Road about 0.5 km west of Sheepshead Island (M. Lupulescu, New York State Geological Survey, written communication, 2019).

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Appendix 1. Representative Photographs of Map Units From the Crown Point Quadrangle

The appendix contains representative photographs (figs. 1.1–1.25) of map units from the Crown Point Quadrangle, Essex County, New York, and Addison County, Vermont. Photographs were taken during fieldwork.



Figure 1.1. Photograph of the Glens Falls Limestone (map unit Ogf) showing fossil hash. Pencil is shown for scale. Photograph by Randall C. Orndorff, U.S. Geological Survey.



Figure 1.2. Photograph of the Crown Point Limestone (map unit Ocp) showing large gastropods, genus *Maclurites*. Pencil is shown for scale. Photograph by Randall C. Orndorff, U.S. Geological Survey.



Figure 1.3. Photograph of the Providence Island Dolomite (map unit Opi) showing dolostone breccia. Rock hammer is shown for scale. Photograph by Randall C. Orndorff, U.S. Geological Survey.



Figure 1.4. Photograph of the Providence Island Dolomite (map unit Opi) showing preserved vertical trace-fossil burrows. Pencil is shown for scale. Photograph by Randall C. Orndorff, U.S. Geological Survey.



Figure 1.5. Photograph of the Whitehall Formation (map unit O€w) showing butcher-block weathering in tan weathering dolostone with black chert and white quartz veins. Rock hammer is shown for scale. Photograph by Gregory J. Walsh, U.S. Geological Survey.



Figure 1.6. Photograph of the Cutting Dolomite (map unit Ocu) showing crossbedded dolostone. Rock hammer is shown for scale. Photograph by Randall C. Orndorff, U.S. Geological Survey.

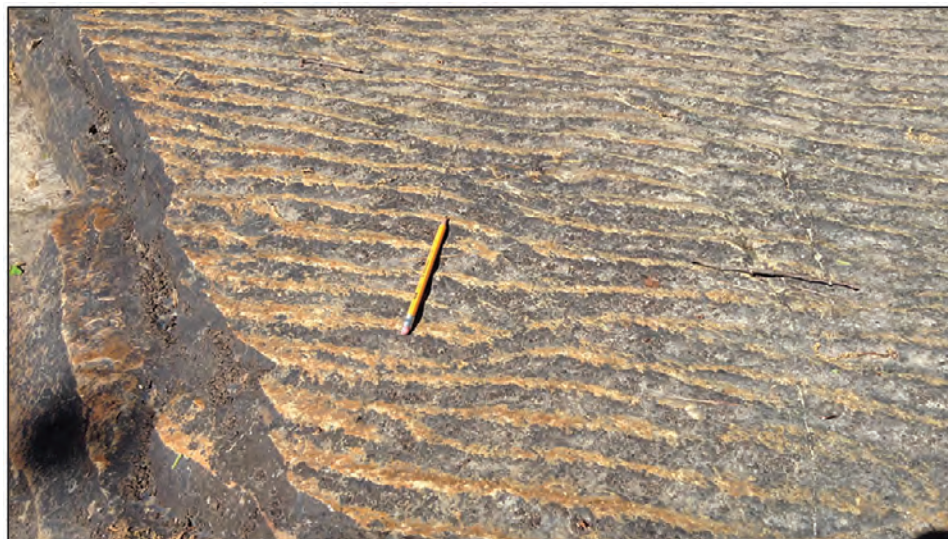


Figure 1.7. Photograph of the Potsdam Sandstone (map unit Cpt) showing ripple marks. Pencil is shown for scale. Photograph by Gregory J. Walsh, U.S. Geological Survey.



Figure 1.8. Photograph of a small field sample of olive-green, aphanitic, mafic dike (map unit Zd, right) in contact with migmatitic biotite gneiss (map unit Ybg, left). Photograph by Gregory J. Walsh, U.S. Geological Survey.



Figure 1.9. Photograph of light-pink leucogranite gneiss (map unit Ylg) of the Lyon Mountain Granite Gneiss with layers of variably assimilated dark amphibolite (map unit Ya) and one quartz vein (Qz) with a pegmatite margin. Rock hammer is shown for scale. Photograph by Gregory J. Walsh, U.S. Geological Survey.



Figure 1.10. Photograph of light-gray leucogranite gneiss (map unit Ylg) of the Lyon Mountain Granite Gneiss intruded into a partially assimilated xenolith of dark amphibolite (map unit Ya). Rock hammer is shown for scale. Photograph by Gregory J. Walsh, U.S. Geological Survey.



Figure 1.11. Photograph of light-pink leucogranite gneiss of the Lyon Mountain Granite Gneiss (map unit Ylgg) with abundant millimeter-scale red garnets on the south slopes of Buck Mountain. Pen is shown for scale. Photograph by Gregory J. Walsh, U.S. Geological Survey.



Figure 1.12. Photographs of transitional migmatitic paragneiss (map unit Ylgt). Photograph A (top) shows a greater percentage of pink leucogranite with some gray paragneiss exposed on the left side. Photograph B (bottom) contains a greater percentage of paragneiss. Compass is shown for scale. Photographs by Gregory J. Walsh, U.S. Geological Survey.

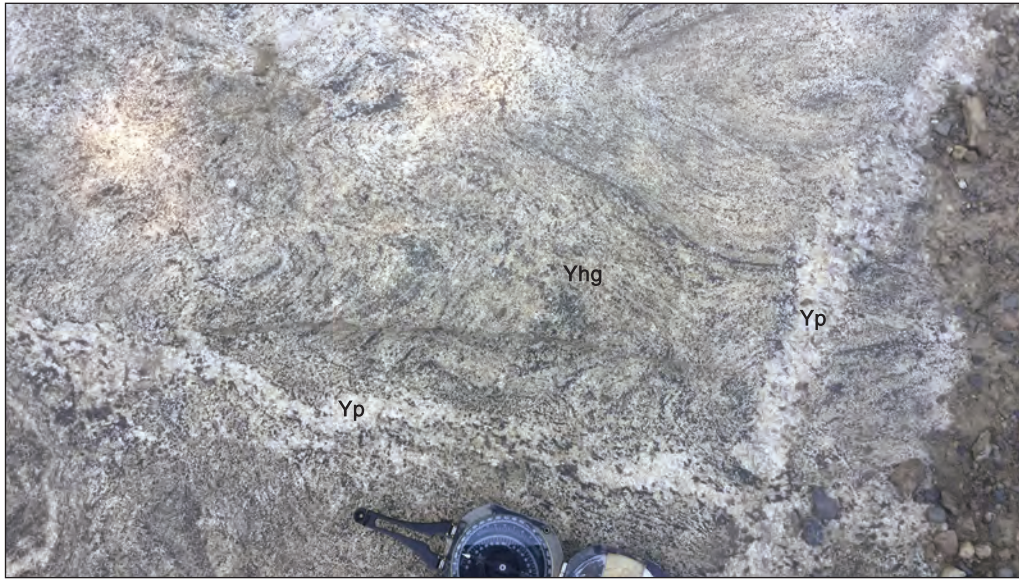


Figure 1.13. Photograph of clinopyroxene-hornblende granitic gneiss (map unit Yhg) with two crosscutting pegmatites (map unit Yp). Compass is shown for scale. Photograph by Gregory J. Walsh, U.S. Geological Survey.



Figure 1.14. Photographs of granitic augen gneiss (map unit Yggn) showing two examples (*A* and *B*) with penetrative foliation and deformed augen of K-feldspar. Marker and pen are shown for scale. Photographs by Gregory J. Walsh, U.S. Geological Survey.



Figure 1.15. Photograph of migmatite gneiss with pegmatite (map unit Ymig) showing abundant reddish garnet. Rock hammer is shown for scale. Photograph by Gregory J. Walsh, U.S. Geological Survey.

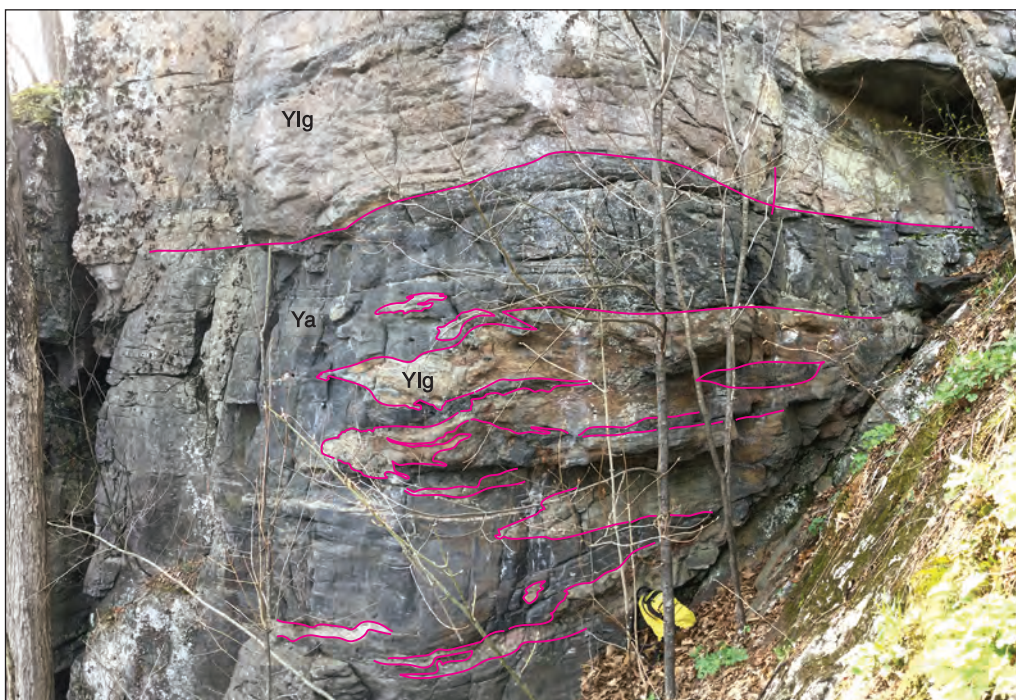


Figure 1.16. Photograph of amphibolite gneiss member (map unit Ya) showing sills and dikes of lighter-colored map unit Ylg (outlined in pink) subparallel to the subhorizontal S2 gneissosity on the east side of Miller Mountain. Yellow backpack is shown for scale. Photograph by Gregory J. Walsh, U.S. Geological Survey.



Figure 1.17. Photograph of amphibolite gneiss member (map unit Ya) showing well-developed foliation parallel to the pencil and layer-parallel leucosomes. Pencil is shown for scale. Photograph by Gregory J. Walsh, U.S. Geological Survey.



Figure 1.18. Photograph of amphibolite, marble, and calc-silicate gneiss member (map unit Yma) showing marble (Ym member, tan) in the foreground and amphibolite (Ya member, dark green) in the background exposed at a gravel pit in the Street Road delta. Compass is shown for scale. Photograph by Gregory J. Walsh, U.S. Geological Survey.

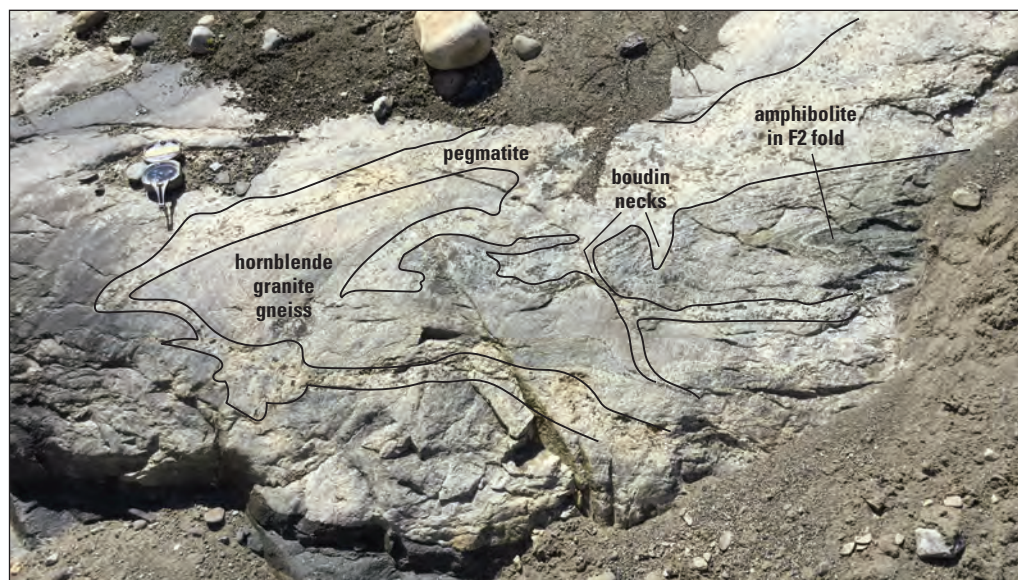


Figure 1.19. Photograph of amphibolite, marble, and calc-silicate gneiss member (map unit Yma) showing migmatitic amphibolite with hornblende granite gneiss in a recumbent isoclinal F2 fold that is exposed at a gravel pit in the Street Road delta. The crosscutting pegmatite occurs as irregular dikes and in D4 boudin necks. Compass is shown for scale. Photograph by Gregory J. Walsh, U.S. Geological Survey.



Figure 1.20. Photograph of marble and calc-silicate gneiss member (map unit Ym) showing coarse-grained marble with graphite flakes at a prospect on Mine Hill. Pencil is shown for scale. Photograph by Gregory J. Walsh, U.S. Geological Survey.

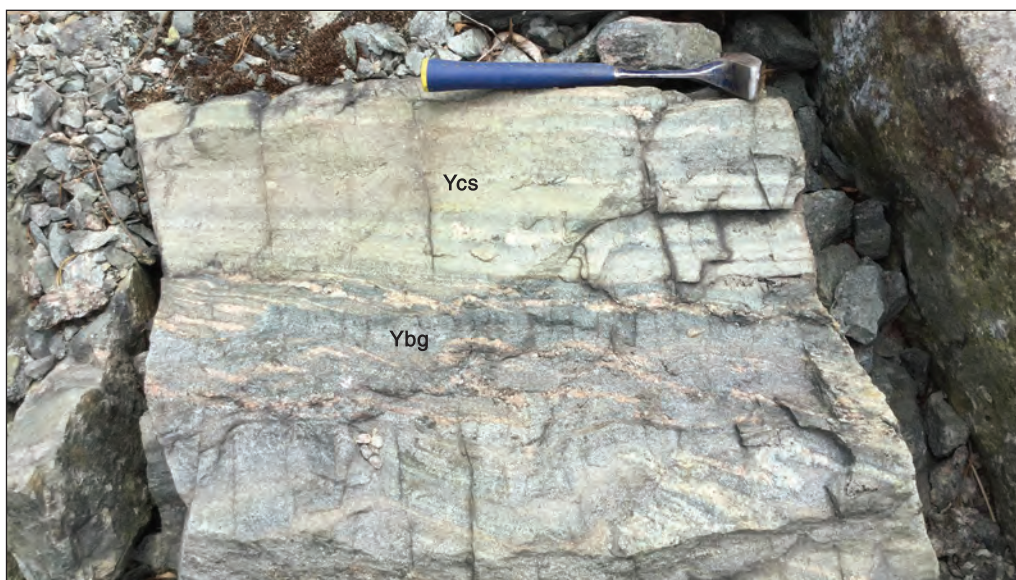


Figure 1.21. Photograph of light-green epidote-tremolite-quartz-diopside granofels layer (map unit Ycs) under a rock hammer within migmatitic biotite gneiss member (map unit Ybg). Photograph by Gregory J. Walsh, U.S. Geological Survey.



Figure 1.22. Photograph of migmatitic biotite gneiss member (map unit Ybg). Compass is shown for scale. Photograph by Gregory J. Walsh, U.S. Geological Survey.



Figure 1.23. Photograph of migmatitic biotite gneiss member (map unit Ybg). Compass is shown for scale. Photograph by Gregory J. Walsh, U.S. Geological Survey.



Figure 1.24. Photograph of migmatitic biotite gneiss member (map unit Ybgg) with garnet and quartz-sillimanite nodules oriented subparallel to F3 fold axes (plunging into the page). Pen is shown for scale. Photograph by Gregory J. Walsh, U.S. Geological Survey.



Figure 1.25. Photograph of garnet-sillimanite gneiss member (map unit YSi) showing quartz-rich layers and abundant garnet. The unit is rusty weathering due to the presence of sulfide minerals. Compass is shown for scale. Photograph by Gregory J. Walsh, U.S. Geological Survey.

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