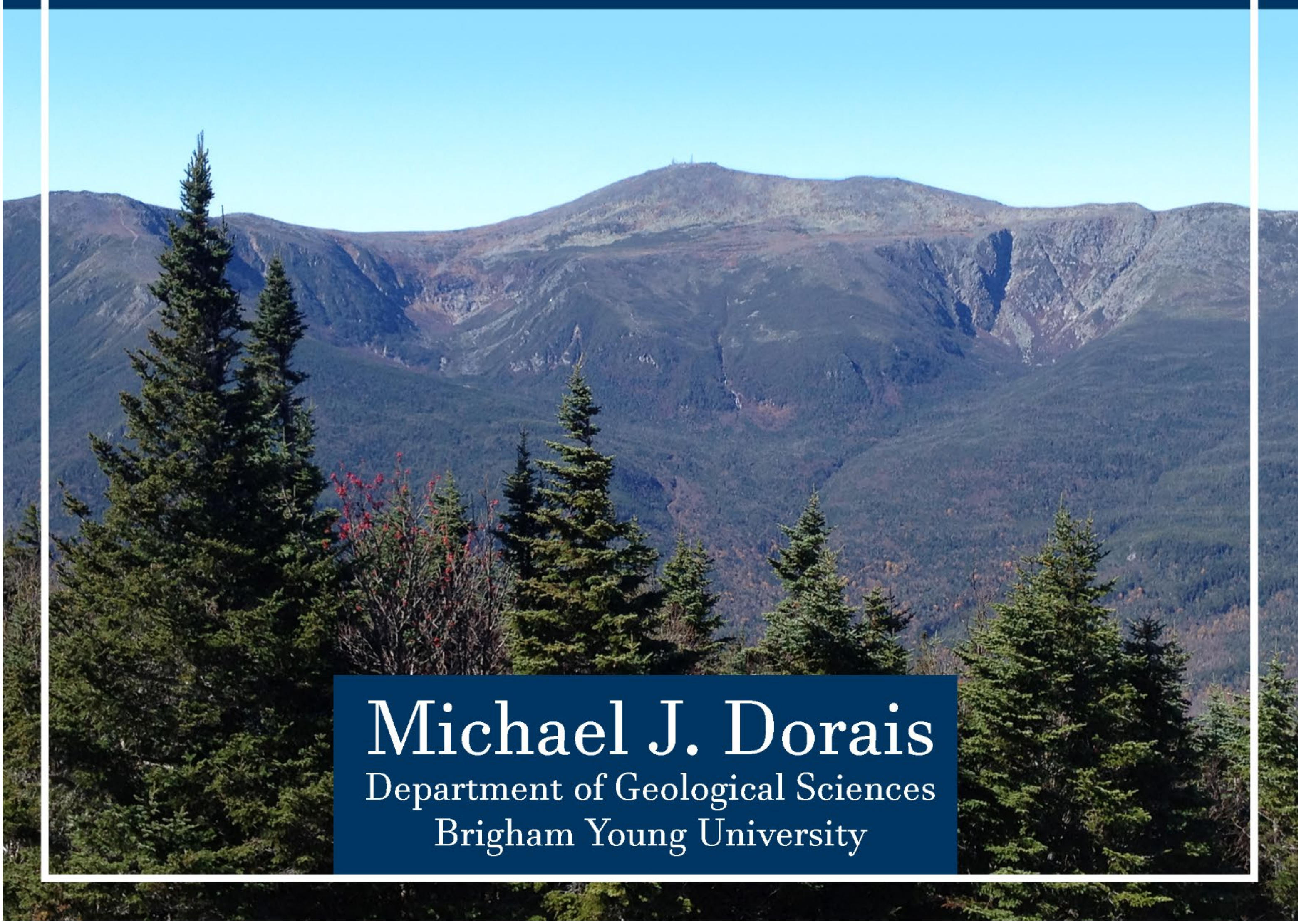
A Field Guide to the Geology of Northern New England



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Front Cover: View of Mount Washington, NH from Wildcat Mountain. The bedrocks consist of Paleozoic metasediments that were shed into the Central Maine Trough after the Taconic Orogeny and metamorphosed during the Acadian Orogeny. Two pre-Laurentide ice sheet cirques are visible, Tuckerman Ravine to the left and Huntington Ravine to the right.

Title Page: Author on Bondcliff, outcrops of the Mt. Lafayette Granite Porphyry, White Mountain Batholith.

Back Cover Top: View of the Northern Presidential Range, NH, from Mount Clay. The bedrocks are Paleozoic metasediments that were shed into the Central Maine Trough after the Taconic Orogeny and metamorphosed during the Acadian Orogeny. The Great Gulf, the glacial valley extending from bottom center of the photo to the right, was carved by an alpine glacier prior to the region having been covered by the Laurentian ice sheet. From left to right: Mount Jefferson, Mount Adams, and Mount Madison. Spaulding Lake, a small glacial tarn, is visible on the floor of the Great Gulf.

Back Cover Bottom: View to Baxter Peak from Pamola Peak, Mount Katahdin, Maine. The mountain consists of the ~ 407 Ma Katahdin Granite.

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Cover design by Jessica Parker

To Ann: The bride of my youth, the love of my life.

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AUTHOR'S PREFACE

Northern New England hosts a wonderful variety of rock types, each with its own tale to tell about its origin. The geologic history is complex, with rocks ranging in age from over a billion years old to as young as 100 million years old. In between these two ages occurred multiple collisional events where island arcs, microcontinents and eventually, the Gondwanan continent collided with the ancestral North America continent called Laurentia. Each of these collisions caused a mountain range to form along the collision zone. This process saw the demise of oceans as the ocean basins between Laurentia and the colliding terrane subducted, consuming oceanic crust. Also occurring between those ages are rifting events where the amalgamated continents broke apart, creating new ocean basins and new oceans. Hence, oceans come and go through geologic time. The geology of northern New England preserves a fascinating record of these collisions and rifting events.

Like a good detective story, much of the enjoyment of deciphering geologic history is found in the process of seeking out clues, analyzing the evidence, and trying to make deductions about past events. This guide book is intended to provide that evidence of the geology of northern New England for the non-specialist. We will visit some of the classic outcrops of northern New England, revealing the clues from the rocks and what these clues say about the larger story of crustal evolution of the northern Appalachian Mountains.

My fellow geologists might review the list of stops in this guide and correctly think that there are many other fantastic locations to visit beyond those presented here. I've attempted to pick stops that are easily accessible and on public land. Most are within easy walking distance from parking locations. This criterion excluded many of classic sites. A few field stops however, have such geologic significance that hikes of several miles are required.

I'm grateful to all the geologists who have labored diligently to unravel the secrets of New England geology. Many of the field stops of this book were chosen from field guides published by the New England Intercollegiate Geological Conferences, a fall tradition of 3 days of field trips that have been held annually for over 100 years. Interested readers may purchase the more specialized descriptions of these sites in guidebooks available on the New England Intercollegiate Geological Conference web page. Many of the pre-1990 guides are available in digital format from the University of New Hampshire at http://www.library.unh.edu/digital/category/science-technology.

Readers will notice that this book focuses on specific locations and does not provide wide-scale descriptions of the entire region. Those wanting a more general view to the geology of northern New England may want to read the Roadside Geology books for Vermont & New Hampshire and Maine (Van Diver, 1987; Caldwell, 1998). These books give good overviews of the geology of the region, however, they lack details of specific outcrops such as those found in this text.

For readers whose interest isn't satiated by this book, there are several other books for the non-specialist that are highly recommended. These include:

Braun, D., Braun, R., 2016. Guide to the Geology of Mount Desert Island and Acadia National Park. North Atlantic Books, Berkeley California, 205 pp.

Caldwell, D.W., 1998. Roadside Geology of Maine, Mountain Press Publishing Company, 317 pp.
Eusden, J.D., 2010. The Presidential Range: Its Geologic History and Plate Tectonics. Durand Press, 62 pp with 1 map.
Eusden, J.D., Thompson, W.B., Fowler, B.K., Davis, P.T., Bothner, W.A., Boisvert, R.A., and Creasy, J.W., 2013. The Geology of New Hampshire's White Mountains. Durand Press, 175 pp.
Gilman, R.A., Chapman, C.A., Lowell, T.V., and Borns, H.W. Jr., 1988. The Geology of Mount Desert Island. Maine Geological Survey, 50 pp with 2 maps.
Hussey, A.M. II, 2000. The Geological Story of Ogunquit, Maine. The Village Press, 33 pp.
Hussey, A.M. II, 2015. A Guide to the Geology of Southwestern Maine. Peter E. Randall Publisher, Portsmouth, NH, 228 pp.
Long, J., 2005. Stepping Stones Across New Hampshire: A Geological History of the Belknap Mountains. Peter E. Randall, Publisher. 80 pp with 1 map.
Rankin, D.W., and Caldwell, D.W., 2010. A Guide to the Geology of Baxter State Park and Katahdin. Maine Geological Survey, 80 pp with 2 maps.
Van Diver, B.B., 1987. Roadside Geology of Vermont and New Hampshire. Mountain Press Publishing Company, 230 pp.

Other good sources of field locations are available on the web page of the Maine Geological Survey that posts a Geologic Site of the Month. The following web site has an interactive map that shows the locations of all these sites:
http://maine.maps.arcgis.com/apps/webappviewer/index.html?id=27f8d4dc1a47416291203683137a3a0b&center=-69.684,43.919&level=11.

The Vermont Geological Survey web site also gives details on interesting sites. The Department of Resources and Economic Development of New Hampshire published geologic maps accompanied by descriptions for the layperson that are available through their web page.

Readers who want to drink deeply in the geology of northern New England should review the geologic maps of Vermont, New Hampshire, and Maine. These are:

Lyons, J.B., Bothner, W.A., Moench, R.H., and Thompson, J.B., Jr., 1997, Bedrock Geologic Map of New Hampshire: Reston, VA, U.S. Geological Survey Special Map, 1:250,000, 2 sheets.
Osberg, Philip H., Hussey, Arthur M., II, and Boone, Gary M. (editors), 1985, Bedrock geologic map of Maine: Maine Geological Survey, 1 plate, correlation chart, tectonic inset map, metamorphic inset map, color geologic map, cross sections, scale 1:500,000.
Ratcliffe, N.M., Stanley, R.S., Gale, M.H., Thompson, P.J., and Walsh, G.J., 2011, Bedrock geologic map of Vermont: U.S. Geological Survey Scientific Investigations Map 3184, 3 sheets, scale 1:100,000.

Rocks similar to those of northern New England extend into Maritime Canada. Two excellent field guide books on neighboring Nova Scotia and Newfoundland are by Hild (2012) and Hild and Barr (2015). These guide books served as inspiration for this guide and I'm pleased to acknowledge following their excellent format.

Hild, M.H., 2012. Geology of Newfoundland: Touring through time at 48 scenic sites. Boulder Publications, Portugal Cove – St. Philip's, Newfoundland. 256 pp.

Hild, M.H., and Barr, S.M., 2015. Geology of Nova Scotia: Touring through time at 48 scenic sites. Publications, Portugal Cove – St. Philip's, Newfoundland. 267 pp.

While the metric system is exclusively used in the sciences, the reader will note that I used the Imperial system for this book. This choice was made to facilitate the use of the book by the lay person with an interest in geology who may not be as comfortable with the metric system. Finally, since this is an online book, it can be edited as need arises. Please send suggestions to me at dorais@byu.edu.

I gratefully acknowledge the assistance of Amy Jeppson, Keili Kwong, and Forrest Strech in producing many of the illustrations used in this publication, and to Eric Christiansen for permission to use some of the illustrations from his book Dynamic Earth: An Introduction to Physical Geology. I'm grateful to Dyk Eusen of Bates College for helpful suggestions on the Maine portion of this book. Dyk has written two excellent books on the geology of the White Mountains that I highly recommend. Much thanks also to Daunine Beck and Kathryn Tucker for assistance with formatting the book and to Jessica Parker for the cover design. Finally, I'm very grateful to my sisters, Carol and Joanne, for years of logistical support during my many trips to NH. We've had a blast hiking the Whites and visiting our secret swimming locations.

Side Note: The reader might wonder how a geologist from Brigham Young University in Utah developed such an interest in New England geology. I was born and raised in New Hampshire. Even though I attended schools out west for my BS and MS degrees, and in Georgia for my PhD, I never lost the strong attachment to northern New England. Prior to joining the faculty at BYU, I was at Indiana University for a dozen years. I'd take my family to vacation in New Hampshire and while there, started picking up rocks from various localities. Thirty plus years later and I'm still picking them up. This guidebook is an attempt to share this fascination with anyone who is interested.

INTRODUCTION

Geologic Time Scale

The miniscule time span of human life on Earth, individually and even collectively, makes the vast extent of geologic time nearly incomprehensible. The Earth is 4.56 billion years old. The first modern form of humans appeared about 200,000 years ago, the last 0.04% of Earth's history. One can justifiably wonder how we know anything about the previous 99.96% of Earth's history since no one was in existence to record it. Reading the rocks to decipher that history is what makes the science of geology so fascinating.

Throughout this book, you'll see references to the ages of geologic events in the New England Appalachians, usually noted in terms of millions to billions of years. For example, the age of the Concord Granite of New Hampshire is about 380 million years, usually abbreviated as 380 Ma with Ma meaning mega annum or 1 million years. Likewise, the abbreviation Ga means giga annum or 1 billion years.

In the early days of geology as a science, geologists could only determine the age of a geological feature with respect to another. For example, the law of superposition decrees that in undisturbed layers of sedimentary rocks, the oldest rocks are at the bottom with successively younger rocks having been deposited upward. Additionally, a dike or sheet of igneous rock cutting across layers of sedimentary rocks required that the sedimentary rocks are older. These and other field relations are used to determine relative ages, i.e., the sequence of events in rocks, but no constraints on absolute ages could be determined, i.e., ages in years. For example, they could observe that a granitic pluton was intruded and then deformed during a collisional event, but no absolute age constraints can be obtained on either the age of the pluton or when it was deformed. Fortunately, pioneering geologists were quite skilled at determine relative ages of rocks. They also determined that fossils succeed each other in a systematic order in progressively younger rocks and that this order can be identified in other outcrops over large distances. This observation led to the identification of similar sequences of rocks in other locations, allowing correlation of strata based on ages. Eventually, after many decades of work, geologists were able to construct the geologic time scale as we have it today (Figure I-1). But it wasn't until radiometric dating techniques became available that geologists were able to determine absolute ages.

Some elements in nature are radioactive and decay to daughter elements at specific rates that have been accurately measured. Minerals that incorporate a radioactive element in their structures, but not the daughter elements, can be used to determine ages. Uranium (U) and thorium (Th), for example, decay to lead (Pb). When a mineral such as zircon ($ZrSiO_4$) crystallizes, it incorporates small amounts of U and Th but no Pb. Over time, as U and Th decay, Pb builds up in the crystal. Geochronologists are able to accurately measure the amount of the parent elements U and Th, and the daughter element, Pb, and because we know the rate of decay, an age of the zircon crystal can be determined.

Dating minerals such as zircon has placed absolute age constraints on the ages of fossils as well and provided absolute ages for the geologic time scale. For example, volcanic ashes commonly contain zircon crystals. Dating ashes just below and immediately above a dinosaur fossil places upper and lower constraints on age of the dinosaur that lies between the two ash beds.

The result of decades of work by numerous geologists is the Geologic Time Scale as displayed in Figure I-1.

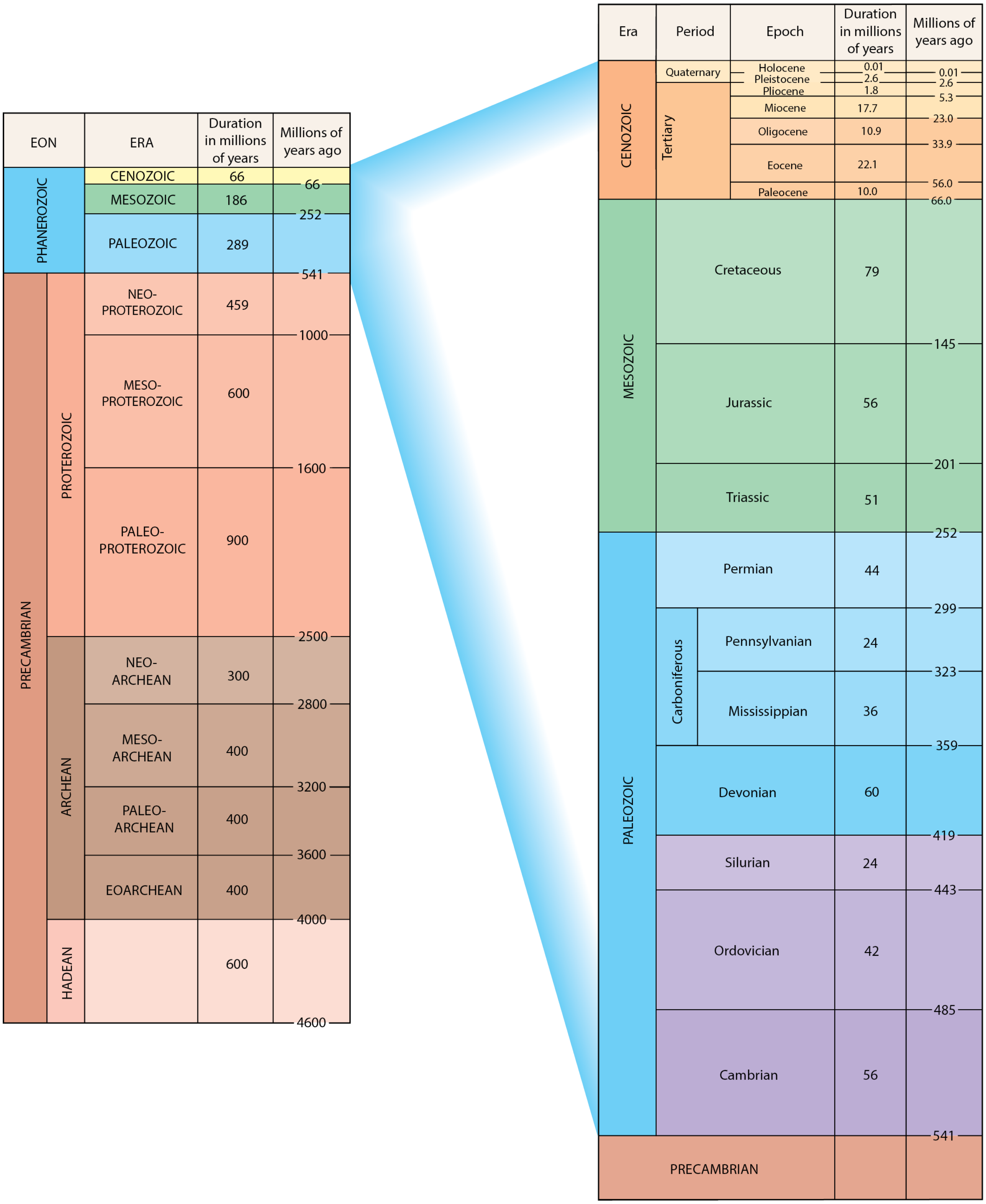


Figure I-1. Geologic time scale (Courtesy of Eric H. Christiansen from Dynamic Earth (2015).

Plate Tectonics

Figure I-2 shows a cross section through the Earth, exposing the Earth's layers. The Earth's core consists of two parts, a central solid, inner core and a molten, outer core. Differential rotation of the two parts of the core generates the Earth's magnetic field. Outside of the core is the Earth's mantle, consisting of ultramafic rocks. The mineralogy of the ultramafic rocks depends on depth; the shallower portions of the mantle consists mainly of olivine and pyroxene. Because of the heat in the interior of the Earth, the mantle is not rigidly solid, rather it can convect and flow. Overriding the mantle is the lithosphere, consisting of solid rocks that extend from the Earth's surface to a depth of 20 miles in the ocean basins and up to 90 miles under the continents.

The lithosphere is not one solid shell that extends around the Earth, rather it is broken into several fragments that geologists refer to as plates (Figure I-3). These plates are less dense than the mantle and are floating at the Earth's surface. Because the underlying mantle convects, lithospheric plates are carried laterally and interact with each other. In locations where the plates are pulled apart, called divergent plate boundaries, the mantle upwells and partially melts, generating basaltic magmas. These magmas ascent into the boundary between the two plates and are best represented by the mid-oceanic ridges. These undersea mountain ranges mark the location of the Earth's youngest crust. Continual divergence of the plates pulls the newly formed crust at the mid-oceanic ridges apart; in essence the new crust is placed on the oceanic basin conveyor belt as successively younger lavas are erupted at the spreading center.

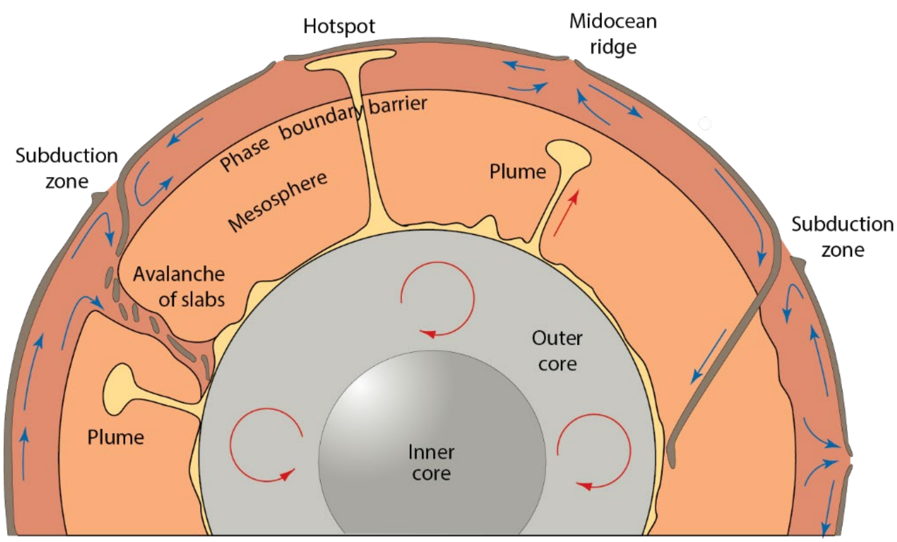


Figure I-2. Cross section of Planet Earth showing the inner and outer core, the lower and upper mantle, and the lithosphere. Also depicted are subduction zones with the lithospheric plate descending below an adjacent plate, and mantle plumes rising through the mantle to impinge on the base of the lithosphere (Courtesy of Eric H. Christiansen from Dynamic Earth (2015).

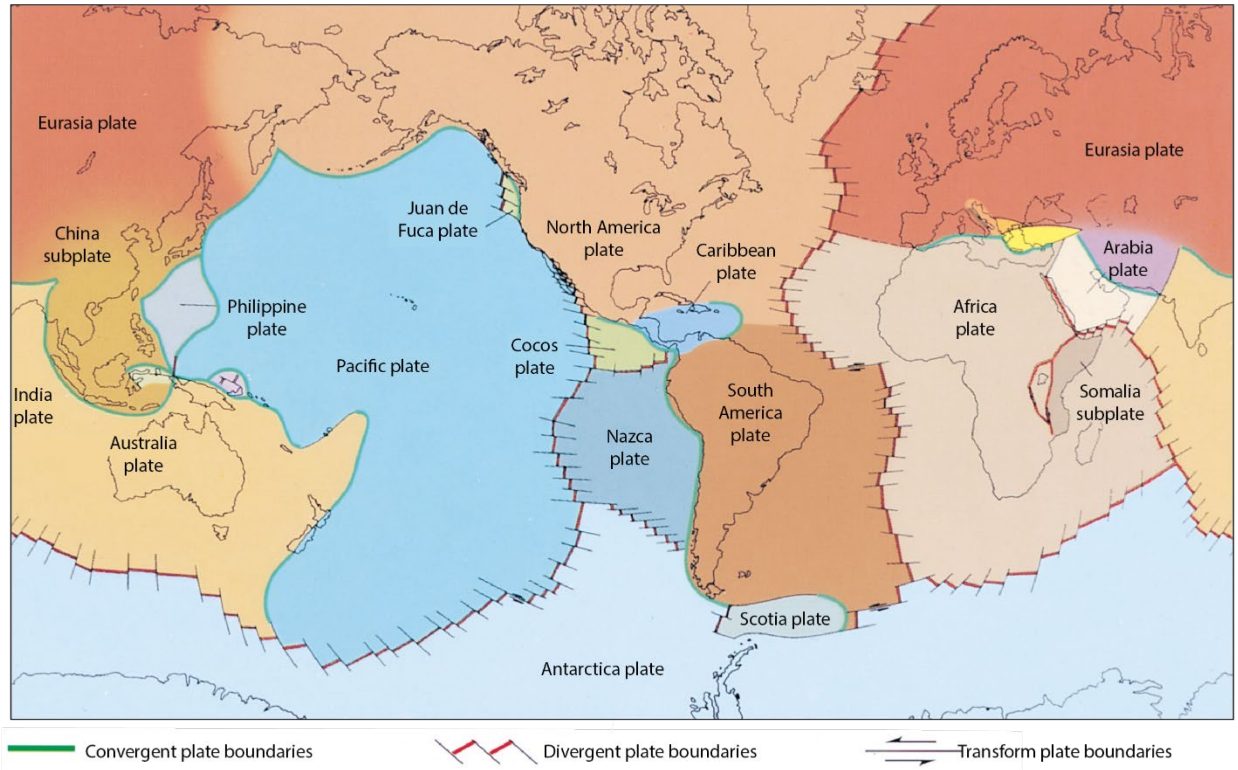


Figure I-3. Map of Planet Earth showing the major tectonic plates and convergent, divergent, and transform (plates sliding past each other) plate boundaries (Courtesy of Eric H. Christiansen from Dynamic Earth (2015).

New oceanic crust cannot be created without older oceanic crust being consumed. Eventually, the oceanic crust migrates away from the spreading center and subducts or slides beneath either another oceanic plate or under a continental plate (Figure I-4). Where an oceanic plate subducts under another oceanic plate, the subduction zone is characterized by a deep, oceanic trench such as the Marianas Trench, and by a chain of overlying volcanoes such as those of the Aleutian Islands and the Philippines (Figure I-4a). These chains of volcanoes are called island arcs. In other locations such as the west coast of South America, the oceanic plate is subducting under continental crust. A chain of volcanoes also developed, but here the volcanic chain is built along the continental margin and is called a continental arc (Figure I-4b).

The oceanic plate acts as a conveyor belt and whatever sits on that belt (or is attached to it) gets carried with the oceanic plate to the subduction zone. Island arcs, continental fragments, and even continents themselves migrate to the subduction zone and because they are less dense than the oceanic crust, these rocks usually do not subduct very deep into the mantle. They clog up the system, shutting down subduction. When a crustal fragment such as an island arc or microcontinent arrives at a subduction zone descending under a continental margin, a collision occurs, causing an orogeny or mountain building event. Just as a car may collide with a stationary car at a stop sign, so will a continent, microcontinent, or island arc collide with a "stationary" continent causing deformation of the rocks involved in the collision. The classic example of continent – continent collision is the Himalaya that resulted from the collision of India with the Asian continent (Figure I-4c). Newly added belts of rock from these collisions are called terranes.

The New England Appalachians experienced multiple collisions between the ancient North American continent called Laurentia and island arcs, microcontinents and the Gondwanan continent over a time period of 300 million years, with each collision building its own specific mountain range. Each collision added new terranes to the Laurentian margin. In fact, most of New England consists of terranes formed offshore of Laurentia that were added, one belt after the other, over this 300 million year time span.

Continent to continent collision form supercontinents such as Rodinia that developed a billion years ago and Pangea that formed 300 million years ago. Supercontinents don't last forever; eventually, a super continent rifts apart, and new oceanic crust with a spreading center develops between the diverging continental fragments. This reoccurring cycle of supercontinent formation, rifting and formation of new ocean basins, followed by subduction and additional continental collision, is called the Wilson cycle after J. Tuzo Wilson, one of the pioneers of plate tectonics.

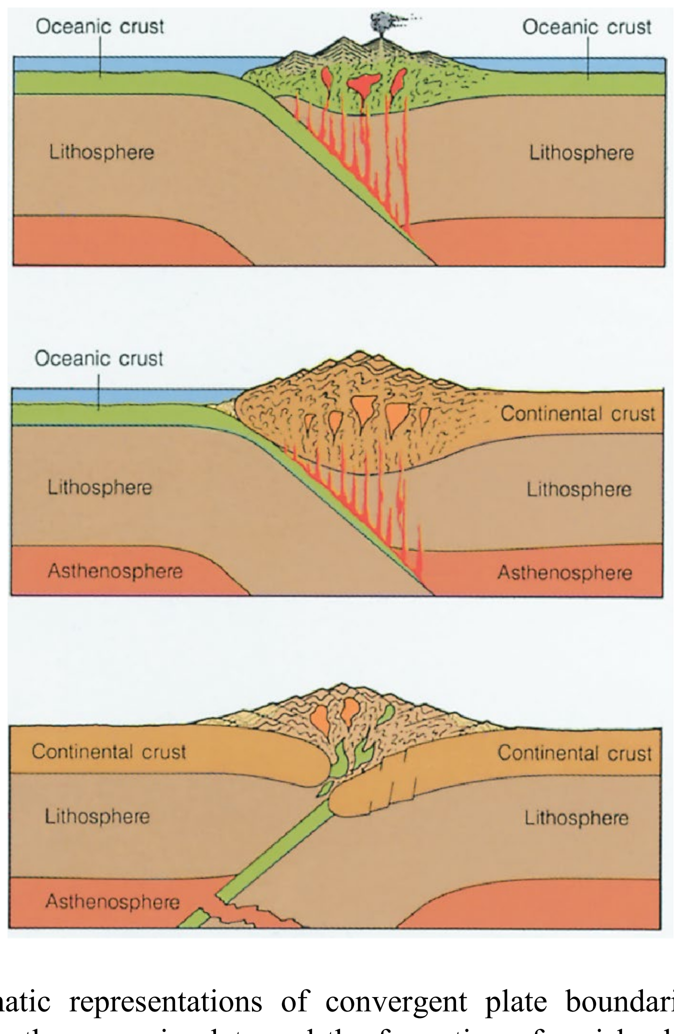


Figure I-4. Schematic representations of convergent plate boundaries. A. Oceanic crust subducting below another oceanic plate and the formation of an island arc volcanic chain. B. Oceanic plate subducting below a continental margin and the formation of a continental arc. C. Continent – continent collision (Courtesy of Eric H. Christiansen from Dynamic Earth (2015).

Rock Cycle

Rocks are classified as igneous, sedimentary or metamorphic. Igneous rocks formed from the cooling of molten rock called magma. A wide range of magma compositions exists, but for our purposes, it is instructive to consider two endmember compositions. Magmas are rich in Fe, Ti, Mg, Ca with moderate Si concentrations are called mafic. These magmas will crystallize minerals indicative of their compositions, forming olivine, pyroxenes and Ca-rich plagioclase feldspar. If the magmas erupt as lavas, the rock is called basalt. If it crystallizes slowly at depth, the rock is a gabbro or diorite. The other endmember is felsic, being richer in Si, Na, and K and poor in Fe, Ti, Mg, and Ca. These magmas form Na-rich and K-rich feldspars and quartz. If erupted, the rock is rhyolite, if crystallized at depth, it is granite. Igneous rocks that crystallized beneath the surface of the Earth are called plutonic rocks. Large plutonic bodies are called plutons.

It may seem pedantic to many readers, but geologists have devised classification schemes for igneous rocks based on ratios of olivine to pyroxene or of plagioclase, potassium feldspar, and quartz. Since northern New England is the host to numerous granitic plutons, Figure I-5 is presented that shows the classification scheme for granitic rocks that is used in this book. The amount of hornblende, biotite and other minerals is ignored and the amounts of quartz, plagioclase feldspar and potassium feldspar are normalized to 100% and are plotted in this triangular diagram to determine the rock name. In this guide, rock names such as syenite, quartz syenite, granite, granodiorite, tonalite, diorite and gabbro will be used.

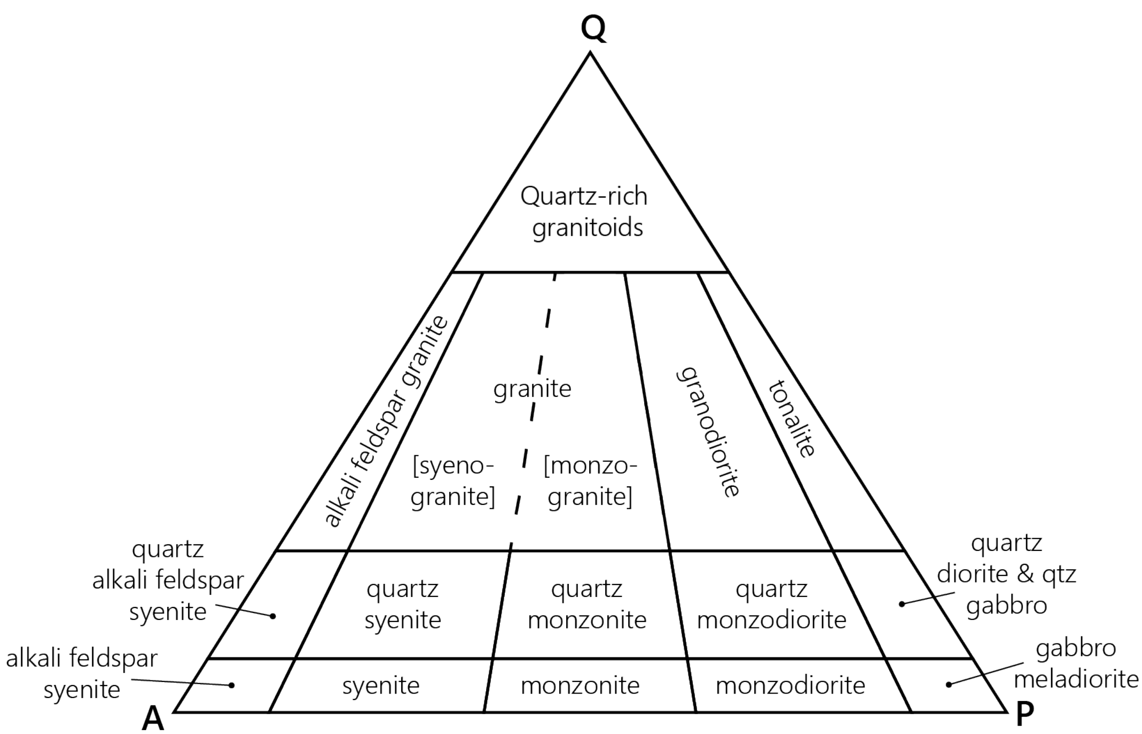


Figure I-5. Classification scheme for plutonic igneous rocks, called the Streckeisen diagram after the author who devised it. Q, A, and P refer to quartz, alkali feldspar, and plagioclase feldspar respectively.

All rocks exposed at the Earth's surface are subjected to the erosional effects of wind, water, and ice, undergoing mechanical weathering which breaks the rocks into fragments of various sizes. Many minerals are not stable at the Earth's surface as well and these undergo chemical weathering to break down to other minerals that are stable at low temperature and pressure and the presence of water. For example, feldspars weather to clays.

Sediments that have not been transported far from their sources tend to be compositionally diverse and contain angular, large particles of a range of grain sizes; these are termed immature sediments. The size of the eroded rock fragments diminishes with distance from their sources because rock fragments are abraded and rounded during transportation in streams. Many of the original minerals in the source rock are also chemical weathered and altered. Well sorted, well rounded sediments that contain stable minerals such as quartz are called mature.

When streams carry sediment into lakes and the ocean, the deceleration of stream velocity allows the sedimentary particles to settle, depositing the sediment as layers. Burial, compaction, and cementation turns the sediment into rock.

Other sedimentary rocks consist of chemical precipitates. Various microscopic marine animals secrete calcium carbonate to form shells. As the animals die, the shells sink to the sea floor, accumulating to form thick layers that eventually are lithified. Calcium carbonate can also directly precipitate from ocean water with deposits also lithifying to limestone.

Both igneous and sedimentary rock can be metamorphosed, particularly as the rocks are involved in collisions at convergent plate margins. Northern New England is blessed with abundant metamorphic rocks that formed as island arcs, microcontinents and the Gondwanan continent collided with ancestral North America. These collisions subjected the rocks to different temperature (T) and pressure (P) conditions, causing the original minerals in the rock to react to form other minerals that were stable under the new T-P conditions. Geologists classify metamorphic rocks based on common mineral assemblages that formed under similar pressure – temperature conditions into distinct fields in pressure – temperature space (Figure I-6). These fields are termed metamorphic facies. Rocks from around the world that have the same composition and metamorphosed to the same pressures – temperatures will have the same mineral assemblage. For example, basalts are common in many mountain belts and typically are metamorphosed to a mineral assemblage dominated by an amphibole called hornblende and plagioclase feldspar. This assemblage is characteristic of the amphibolite facies.

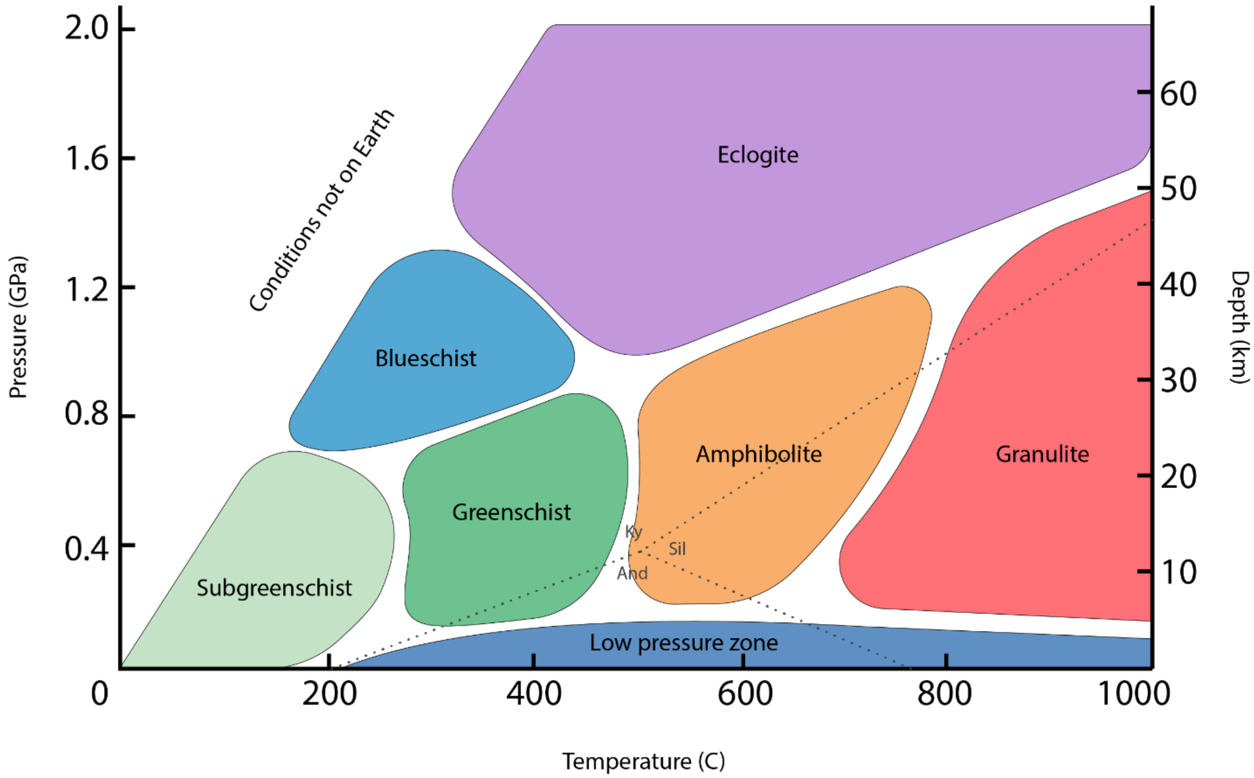


Figure I-6. Pressure – temperature diagram showing the fields for various metamorphic facies.

This field guide will take you to see many metamorphic rocks, some of which attained greenschist facies metamorphism, others were subjected to higher degrees of metamorphism to the amphibolite and granulite facies. Metamorphosed clay-rich sedimentary rocks host a wide variety of metamorphic minerals depending of the pressure – temperature conditions of metamorphism. We'll see rocks with garnet, andalusite, sillimanite, staurolite, etc (greenschist and amphibolite facies). Some rocks of northern New England reached granulite facies conditions and were sufficiently hot that the rocks melted, forming granites. In Vermont, we'll visit high pressure, low temperature rocks of the blueschist facies indicating that the rocks descended down into the subduction zone to experience high pressure conditions.

Faults and Folds

Faults occur in rocks when the rocks break and slide past each other along a fault plane. Several types of faults occur in nature, all of which are found in northern New England. When rocks are pulled apart by tension, a normal fault is produced where one block moves down with respect to the other (Figure I-7A). The down dropped block is called the hanging wall, the lower block below the fault is the footwall. In some cases, extension is sufficient that a graben or rift valley will form with normal faults forming both sides of the valley (Figure I-7D). In contrast, rocks that undergo compression will sometimes fracture with one block moving up over the other. These are reverse faults (Figure I-7B), where the hanging wall moves up over the footwall. Low angle reverse faults are called thrust faults where the hanging wall may move many miles over the lower block (Figure I-7E). We'll see several thrust faults in Vermont and New Hampshire that resulted from collision events. Other faults are strike-slip faults where the blocks simply move sideways past each other (Figure I-7C). Strike-slip faults are either right-lateral (dextral) or left-lateral (sinistral). If the viewer is standing on one block, looking across the fault to the other, and that block moved to the right, the strike-slip fault is right-lateral. In contrast, if the block moved to the left, the fault is left-lateral. The well-known, classic strike –slip fault is the San Andreas fault in California. A similar but more deeply eroded fault is the Norumbega Fault System that we'll visit in Maine.

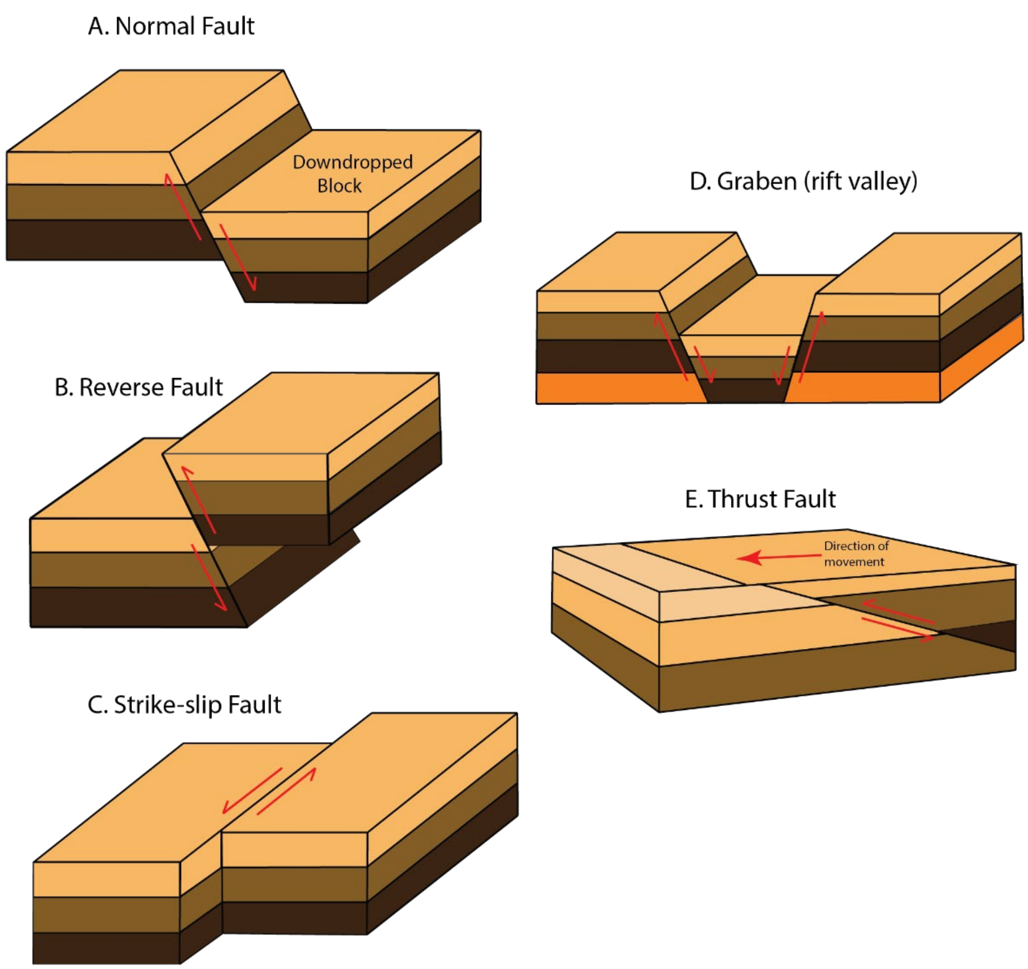


Figure I-7. Schematic illustration of normal, reverse, and strike-slip faults.

At depth where the rocks are hotter than those that undergo faulting, deformational events may deform the rocks by folding. In some cases, the folds may be small (e.g., Figure 7-3), in others, the folds extend for miles (Figure NHS-2). These large folds are called nappes, and were first recognized in the Alps. Eventually, the large folds of the nappes may break, transforming into a trust fault. In New England, geologists tend to use the terms nappe and thrust fault interchangeably. We will visit several thrust faults in Vermont (Stops 6 and 7) and New Hampshire (Stop 14).

OVERVIEW OF THE GEOLOGY OF NORTHERN NEW ENGLAND

The geology of New England is complex with numerous events contributing to the final assembly of rocks that constitute the geology of the region. The large number of names of terranes, microcontinents, super continents, and orogenic events easily overwhelm the beginner in New England geology. Some of the names of these geologic events and the tectonic elements are summarized in Table 1 which serves as quick overview of the discussion presented below.

The age of the Earth is 4.56 billion years. Rocks of that age are not preserved in the Earths' crust; the oldest rocks are about 3.8 billion years old and are found in Precambrian shields of continents such as the Canadian shield of North America shown in Figure O-1. These Archean to Early Proterozoic rocks consist mainly of granitic gneisses that formed during various orogenic events across ~3 billion years. The shield contains several terranes, accreted belts of rocks, that joined to form the oldest, stable portion of North America. Subsequently, younger orogenic events added additional terranes to the continent. These include the Late Proterozoic Grenville Orogeny, the Paleozoic orogens of the Appalachian Mountains, and the Mesozoic to Cenozoic orogens of western North America (Figure O-1).

The rocks of New England are far younger than those of the Canadian Shield, with the oldest rocks in northern New England being only about 1 billion years old. These rocks formed during ancient mountain building events called the Grenville Orogeny. The Grenville Orogeny was a long lived event that formed mountain ranges extending from Labrador to Mexico. It consisting of several separate collisional events with the Rigolet and Ottawan phases occurred between 980 to 1090 billion years ago and the older, Shawinigan and Elzevirian phases extend between 1140 to 1240 million years ago. Each of these events added new belts of rock to the margin of the proto-North American continent (the Late Proterozic belt shown in Figure O-1), and eventually a large supercontinent called Rodinia was formed as ancestral North America called Laurentia and Gondwana joined.

Grenville rocks are present in the basement of western New England, lying beneath the Paleozoic rocks exposed at the surface. The only Grenvillian rocks exposed in northern New England are found in the Green Mountains of Vermont (Stop 1).

Table 1. Summary of Geologic Events that formed Northern New England

AGE	GEOLOGIC EVENT	PLATE TECTONIC EXPLANATION
Jurassic - Cretaceous	Emplacement of the White Mountain Magmatic Suite	Anorogenic magmatism emplaced along structural lineations
Jurassic	Eruption of the Central Atlantic Magmatic Province	Rifting of Pangea and the opening of the Atlantic Ocean basin
Permian	Alleghanian Orogeny	Accretion of Gondwana to modified Laurentia
Late Devonian	Neoacadian Orogeny	Accretion of Meguma to modified Laurantian margin
Early Devonian	Acadian Orogeny	Accretion of Avalonia to modified Laurentian margin
Late Silurian - Early Devonian	Formation of the Central Maine and Connecticut Valley-Gaspe Troughs	Orogenic collapse of the Taconic Mountain Range
Late Silurian	Salinic Orogeny	Accretion of Ganderia to modified Laurentian margin
Mid-Ordovician	Taconic Orogeny	Accretion of the Taconic island arc to the Laurentian margin
Cambrian-Ordovician	Penobscot Orogeny	Amalgamation of Ganderian microplates within the Iapetus Ocean
Latest Neoproterozoic	Iapetan Event	Rifting of Rodinian supercontinent and the opening of the Iapetus Ocean
Late Neoproterozoic	Grenville Orogeny	Assembly of the Rodianian supercontinent

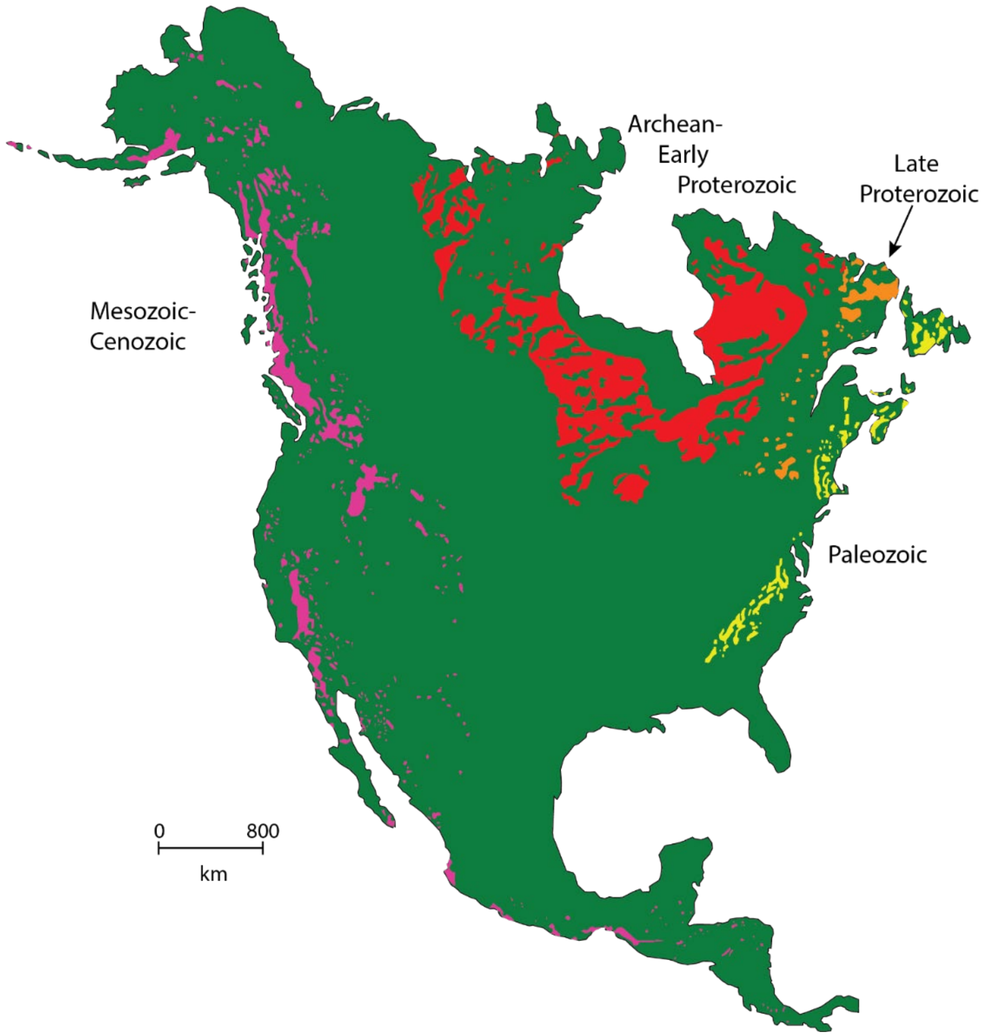


Figure O-1. Generalized map of North America showing the locations of the Archean to Early Proterozoic granitic rocks of the Precambrian Shield (red), the Late Proterozoic plutons of the Grenville Orogeny (orange), the plutons from multiple orogens during the Paleozoic (yellow) and Mesozoic to Cenozoic plutons (purple). Each of these orogenic events added new crust to the North American continent (After Clarke, 1992).

About 550 million years ago, the Rodinian supercontinent began to break apart with the Iapetus Ocean forming between Laurentia and the outboard, rifted continental called Gondwana (Figure O-2). This ocean was named after the Titan Iapetus of Greek mythology who was the father of Atlas for whom the Atlantic Ocean was named. Within the Iapetus Ocean basin lay the Taconic Island Arc, and the Ganderian and Avalonian microcontinents (Figures O-6, O-8)

When the Rodinian supercontinent rifted, the thickness of the continental crust decreased, allowed hot mantle rocks to rise (Figure O-3). The ascent of hot mantle rocks causes partial melting and intrusion and eruption of basaltic magmas along the growing rift (Figure O-4). The thinning of the crust also produced low-lying basins with sediment from the adjacent highlands being shed into these basins (Figure O-3).

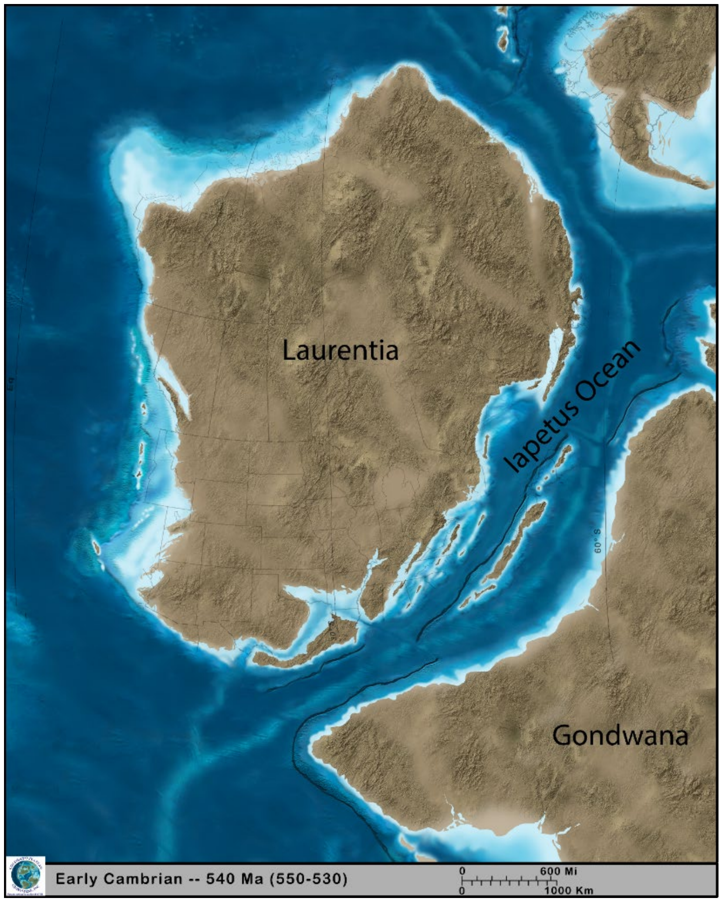


Figure O-2. Rifting of Rhodinia began about 550 million years ago as Gondwana splits from Laurentia and the Iapetus Ocean formed. Image by Ron Blakey.

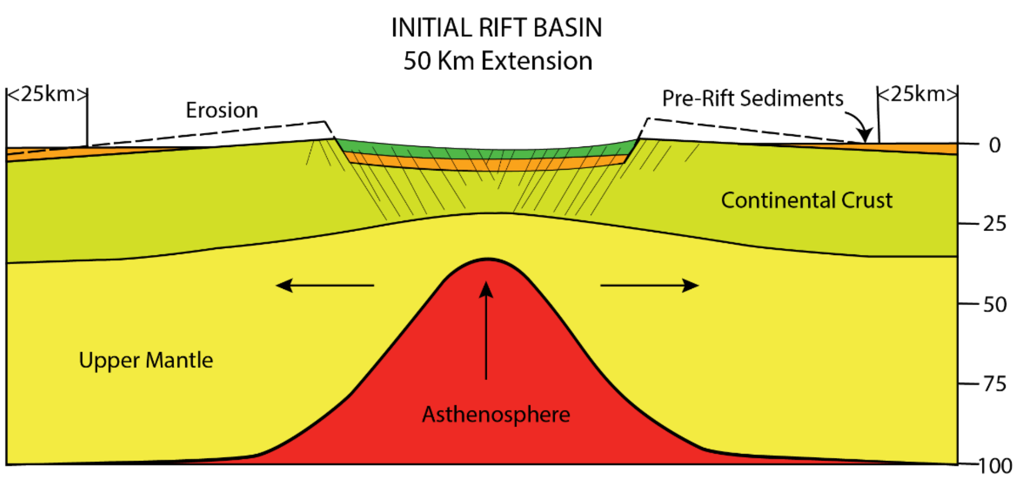


Figure O-3. Cross section showing schematized image of the initial stages of continental rifting. Crustal thinning causes the asthenosphic mantle to rise and a rift basin to form. Immature sediments from the adjacent highlands are deposited in the rift basin (After J. Tarney, Plate Tectonics: Geological Aspects).

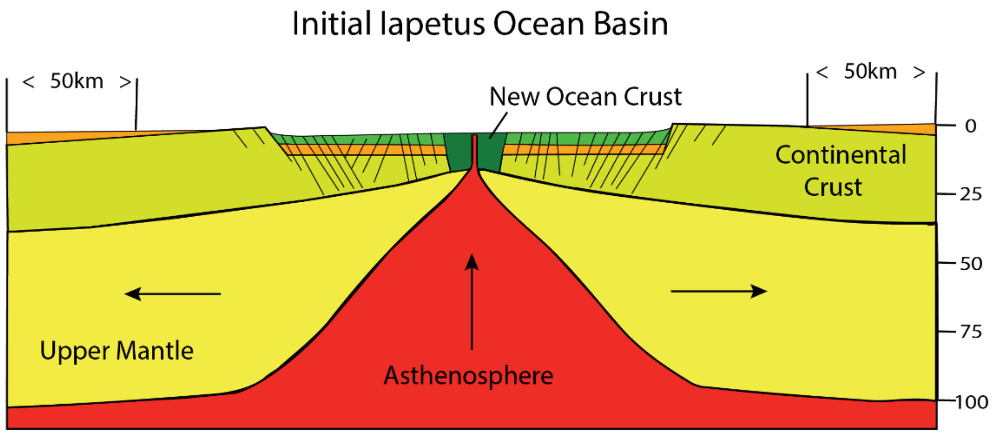


Figure O-4. Continued ascent of the asthenospheric mantle causes partial melting and basaltic magmas intrude into the rift zone. Some of these magmas erupt as large lava flows forming a flood basalt province. Other basaltic magmas remain as dikes, sheets of igneous rock that served as feeders to the lava flows on the surface (After J. Tarney, Plate Tectonics: Geological Aspects).

About 500 million years ago, the continental margin matured and entered into what is called the drift stage, i.e., the eastern margin of Laurentia became a passive margin with an offshore continental shelf and slope (Figure O-5). The current Atlantic coast of North America is a passive margin. The term passive margin signifies that no tectonic activity is occurring along the margin. The margin was simply the site of sediment deposition as weathering eroded the landscape and deposited the sediments seaward along the continental shelf and slope. Some of the sediments carried to the margin along the shore line were more mature than the basin deposits of Stop 3, being quartz-rich (Stop 4).

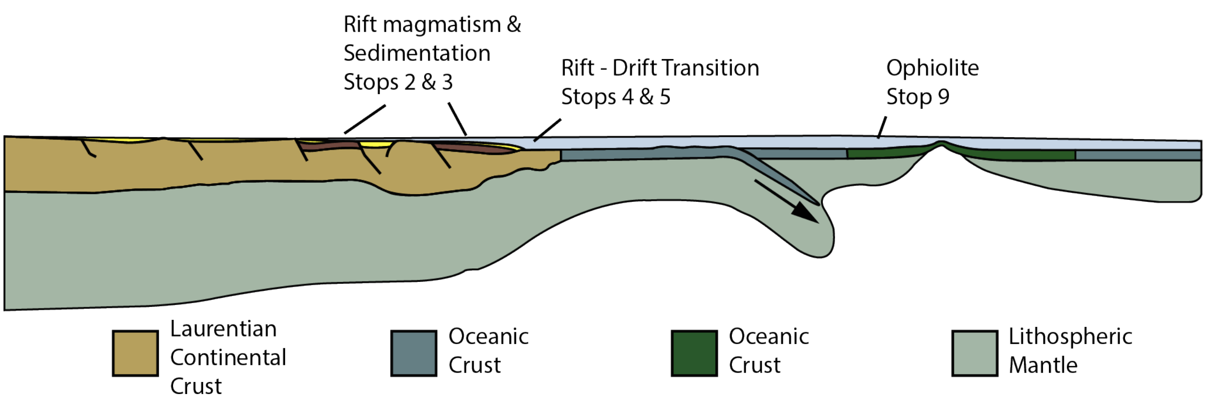


Figure O-5. Geodynamic setting of the northern New England Appalachians during the late Cambrian to early Ordovician (500-470 Ma) showing the original geologic settings of Stops 2,3,4,5 and 9 (After Trembley and Pinet, 2016).

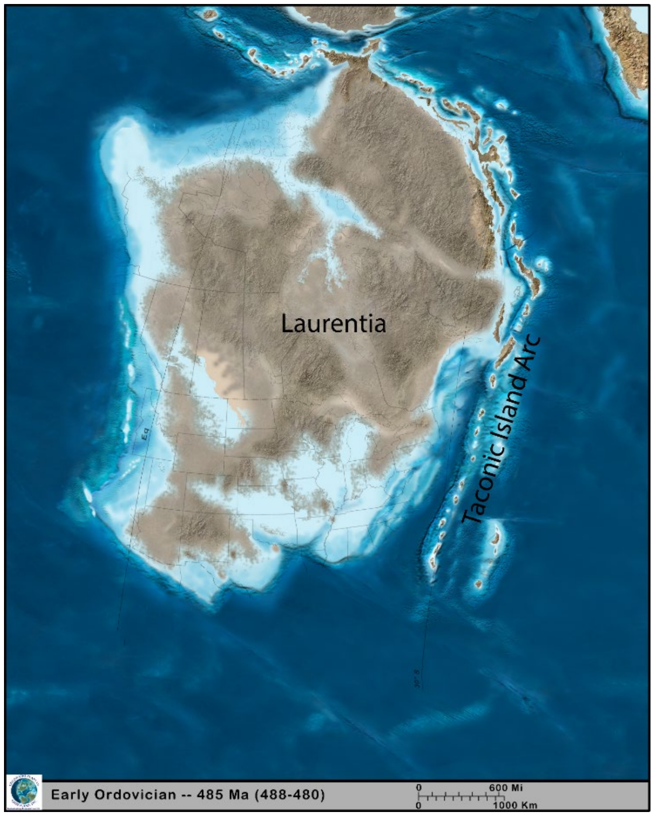


Figure O-6. About 485 million years ago, subduction initiated in the Iapetus Ocean, forming the Taconic Island Arc. The ocean crustal rocks were consumed in the subduction zone and the island arc migrated closer to the Laurentian margin. Image by Ron Blakey.

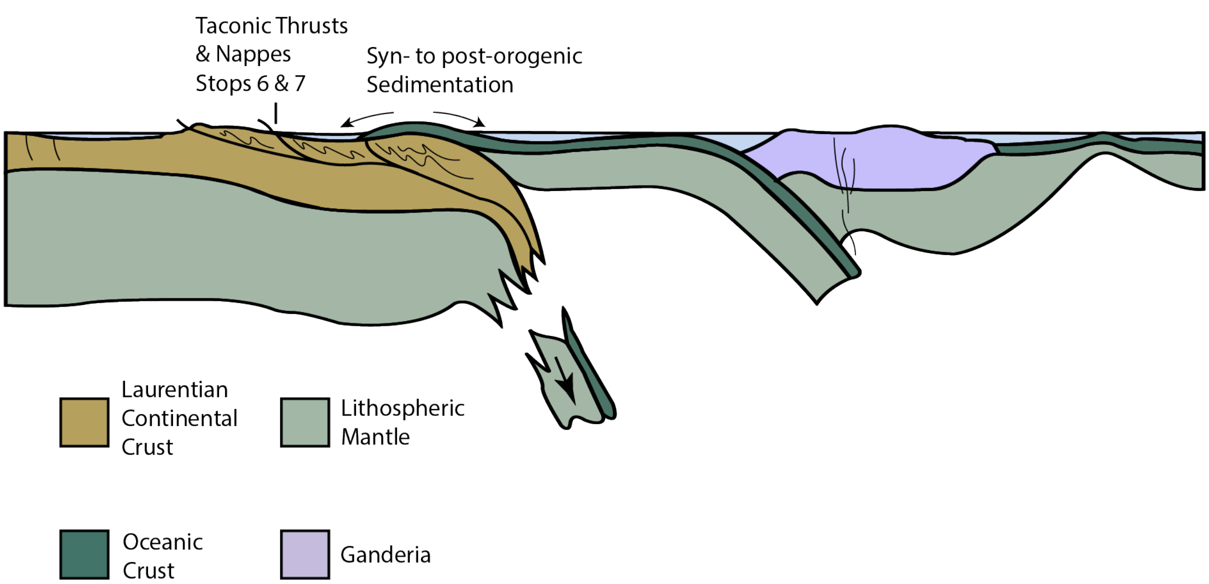


Figure O-7. Geodynamic setting of the northern Appalachians during the middle to late Ordovician showing the thrust nappes of the Taconic Orogeny (After Trembley and Pinet, 2016).

About 480 million years ago when Laurentia resided at tropical latitudes, a subduction zone formed in the Iapetus Ocean off Laurentia's eastern margin (Figure O-5), forming a volcanic arc (Figure O-6). As the ocean basin between Laurentia and the island arc began to be consumed, the island arc migrated closer and closer to the Laurentia, depressing the Laurentian margin. Carbonate reefs formed in this tropical setting, with the Chazy Reef now representing the world's oldest reef complex (Stop 5). Eventually, the edge of Laurentia began to enter the subduction zone and interlaying sediments that were shed off the Laurentian margin into the trough between Laurentia and the island arc were deformed, metamorphosed, and pushed westward over the Laurentian margin (Figure O-7; Stops 6 and 7). The collision caused belts of rock called thrust sheets from deeper settings to be pushed up over younger rocks to the west (Figure I-7E). Some rocks were deeply subducted into the subduction zone, reaching depths of 30 miles or more before being tumbled back up the subduction zone and thrust westward during collision of the island arc (Stop 8). Fragments of deep oceanic crust consisting of mantle rocks were broken and thrust over the Laurentian margin as well (Figure O-7; Stop 9). Eventually, the island arc itself was pushed up over the Laurentian margin (Stop 12). All these features developed during the Taconic Orogeny. Because of the greater buoyancy of continental crust compared to oceanic crust, the continental margin could only subduct so far, eventually clogging the system and causing subduction to cease and the termination of the Taconic Orogeny. Some geologists think that after the accretion of the Taconic island arc, a new subduction zone system developed with oceanic crust being subducted to the west under the Laurentian margin to form a continental arc (Stop 13).

A small microcontinent named Ganderia docked with Laurentia about 430 million years ago during the Salinic Orogeny (Figures O-7, O-8). This docking must have been a relatively gentile collision because there is not much evidence for Salinic metamorphism in New England. But Ganderia forms the basement of much of New Hampshire and Maine.

Once these compressive forces ceased, the mountainous eastern margin of Laurentia was gravitationally unstable and the mountain range began to collapse. This caused basins to form along the continental margin with abundant sediments from the highlands being shed into what is now known as the Connecticut Valley – Gaspé Trough (Stops 10 and 11), the Central Maine Trough (Stops 14, 19, 27, 28, 29; Figure O-9).

Figure O-8. About 430 million years ago, subduction brought the Ganerian microcontinent in collision with the Laurentian margin, causing the Salinic Orogeny. Image by Ron Blakey.

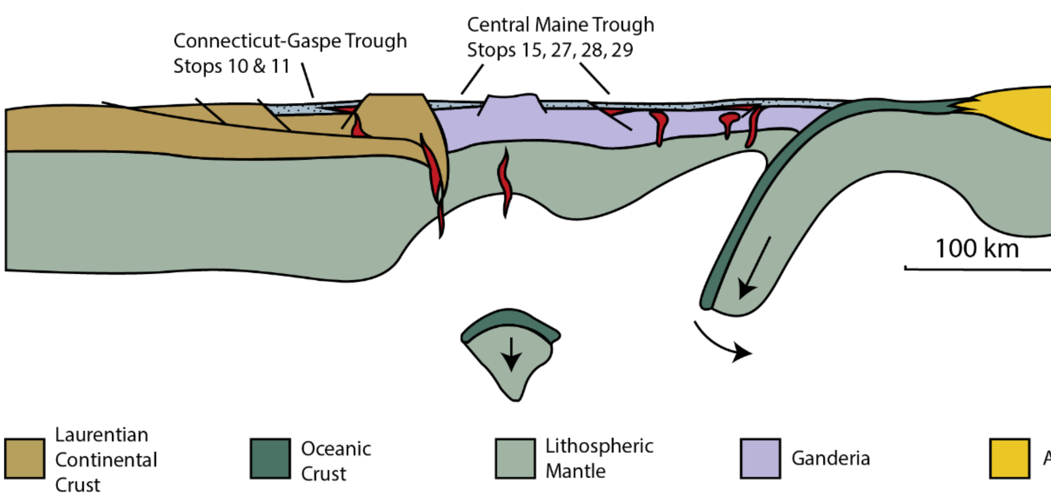


Figure O-9. Geodynamic setting of the northern Appalachians during the late Silurian to early Devonian showing the original locations of Stops 10,11, 15, 27, 28, and 29 (After Trembley and Pinet, 2016).

Like the sediments that were deformed, metamorphosed and thrusted to the west during the Taconic Orogeny, these post-Taconic trough sediments that were deposited in the Connecticut Valley – Gaspe, Central Maine, and Fredericton Troughs were also deformed, metamorphosed and thrusted (Stop 14) during the next orogenic event called the Acadian Orogeny. The Acadian Orogeny was caused by subduction consuming another oceanic basin between Laurentia and an outboard microcontinent called Avalonia (Figures O-10, O-11). The addition of the Avalonian microcontinent to the Laurentian margin produced a mountainous region similar to the Alps or the Himalaya. As continental crust became thickened and depressed deeper into the crust, it eventually heated and caused partial melting. These magmas were squeezed upward and caught in the westward push of the metasediments during the Acadian Orogeny (Stops 15, 16, and 17). In some locations, we observe this high degree of metamorphism where the rocks preserve evidence of partial melting (Stop 15). Accumulation of the melts led to the formation of plutons, large bodies of granitic rock that were emplaced across much of northern New England (Stops 17-19).

The collision of Avalonia with Laurentia marks the complete consumption of the Iapetus Ocean. Eastward of Avalonia (present day coordinates) lay the Rheic Ocean (Figure O-10), named after Rhea, the sister of Iapetus in Greek mythology. This ocean basin contained the microcontinent Meguma and was bordered by the Gondwanan continent on its opposite side.

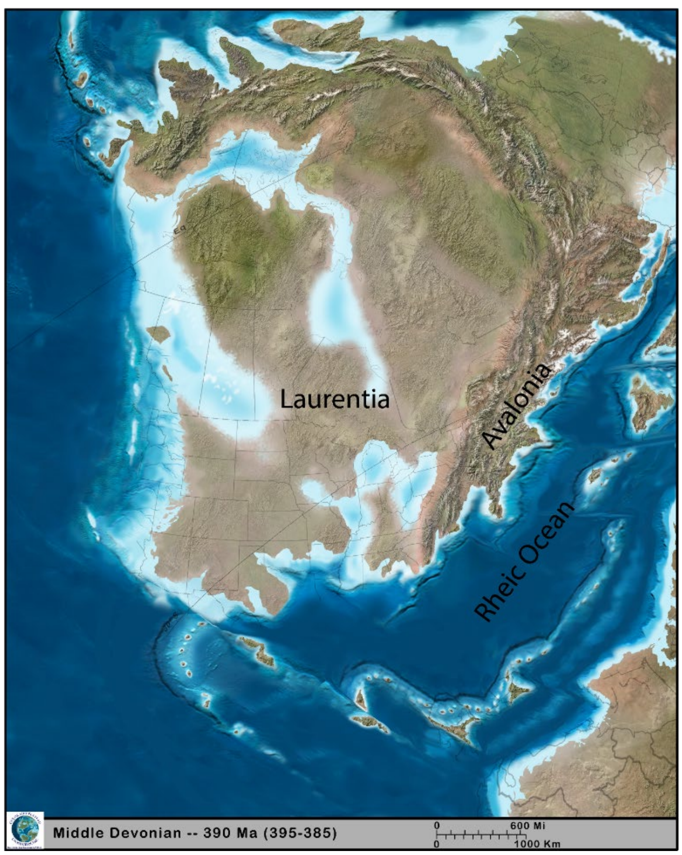


Figure O-10. About 410 to 390 million years ago, the collision of Avalonia with the Laurentian margin caused the Acadian Orogeny. Image by Ron Blakey.

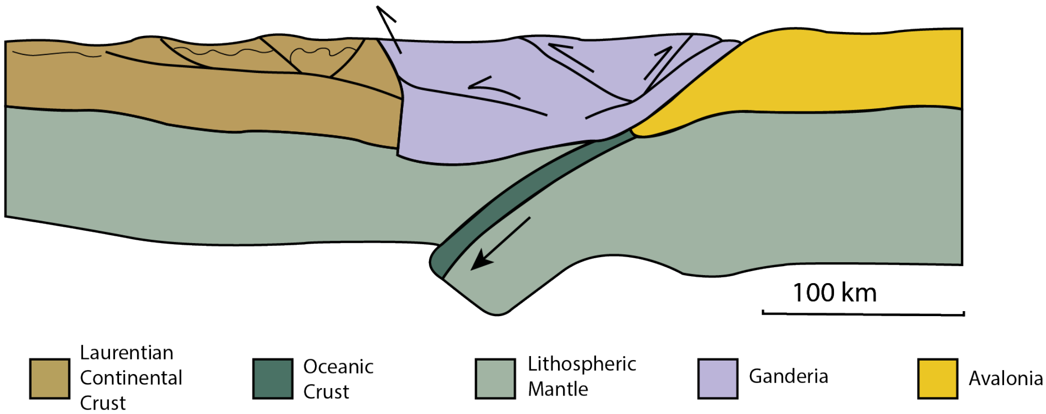


Figure O-11. Geodynamic setting of the northern Appalachians during the Middle Devonian, showing the collision of Avalonia causing the Acadian Orogeny (After Trembley and Pinet, 2016).

The microcontinent Meguma collided with Laurentia about 360 million years ago. No Meguma rocks are exposed in New England, but much of Nova Scotia is underlain by Meguma bedrock. The docking of Meguma caused the Neoacadian Orogeny (Figure O-12).

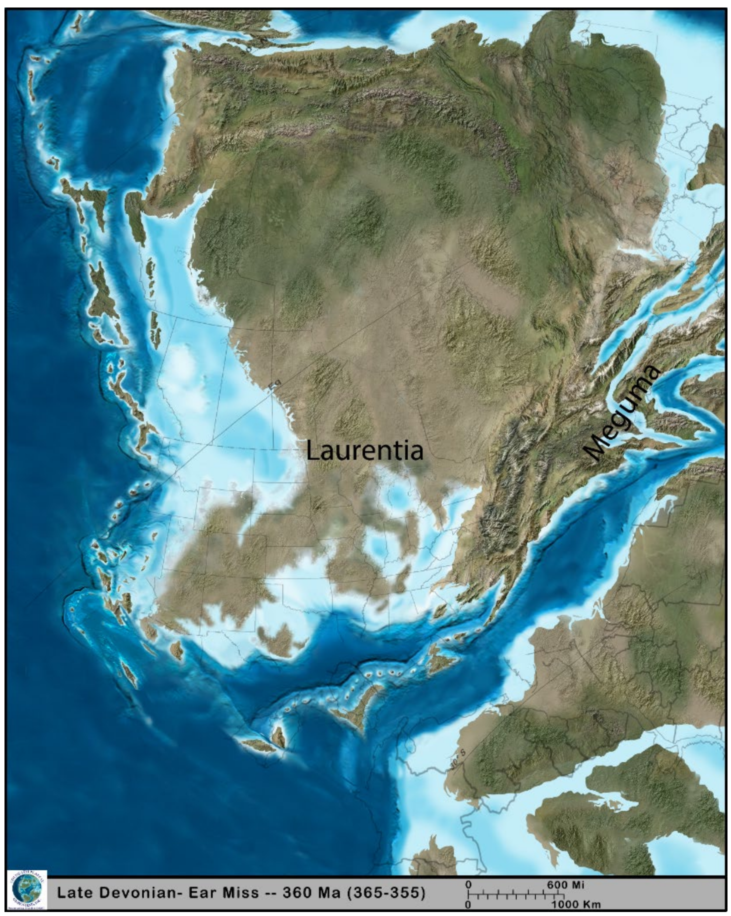


Figure O-12. About 360 million years ago, collision between the microcontinent Meguma and Laurentian caused the Neoacadian Orogeny. Image by Ron Blakey.

Final closure between Laurentia and the continent of Gondwana occurred about 300 million years ago during the Alleghanian Orogeny to form the supercontinent of Pangea (Figure O-13). This collision marked the complete consumption of the Rheic Ocean basin. It was the last orogenic event along the eastern margin of North America, completing the composite Appalachian Mountains. Ganderian rocks, now exposed as the Massabesic Gneiss Complex of southern New Hampshire, partially melted and rose upward at this time (Stop 21).

As Ganderia, Avalonia and Meguma approached and collided with Laurentia, they brought rocks containing the history of events that occurred on the opposite side of the Iapetus and Rheic Oceans. Subduction took place in the eastern portions of the Iapetus and Rheic Oceans just as it did along the Laurentian margin and magmas generated in those subduction zone systems were emplaced in those microcontinents (Stop 35). Fossil assemblages of fauna that lived on the opposite side of these oceans were brought westward. They are recognized as European assemblages that now lie adjacent to native North American assemblages because the microcontinents have been added to the Laurentian crust (Stop 30). Thus the entire Iapetus and Rheic Ocean basins that once were located between these two disparate fossil assemblages has been completely consumed by subduction.

Figure O-13. About 300 million years ago, the oceanic tracks between Laurentia and Gondwana were entirely consumed and the continent – continent collision formed the supercontinent Pangea. Image by Ron Blakey.

The Alleghanian Orogeny was not the last magma-producing event in New England. About 200 million years ago, the supercontinent Pangea began to rift apart (Figure O-14), just like its precursor supercontinent, Rodinia. Pangea thinned and massive amounts of basaltic magma erupted to form one of the Earth's largest flood basalt provinces called the Central Atlantic Magmatic Province. As the Atlantic Ocean basin continued to grow, these volcanic rocks that were once together in a large flood basalt province, were pulled apart to their current locations along the east coast of North America, the northern portions of South America, western Africa and Spain (Figure O-15).

Figure O-14. About 200 million years, Pangea began to rift and the Atlantic Ocean basin began to develop. Massive volumes of basaltic lavas were erupted during this rifting, forming the Central Atlantic Magmatic Province (Figure O-15). Image by Ron Blakey.

No lava flows are preserved in northern New England, but flows are present in the Hartford Basin of Connecticut and along the Bay of Fundy in Nova Scotia. However, feeder dikes to the flows are present in New Hampshire and Maine (Stops 22 and 42).

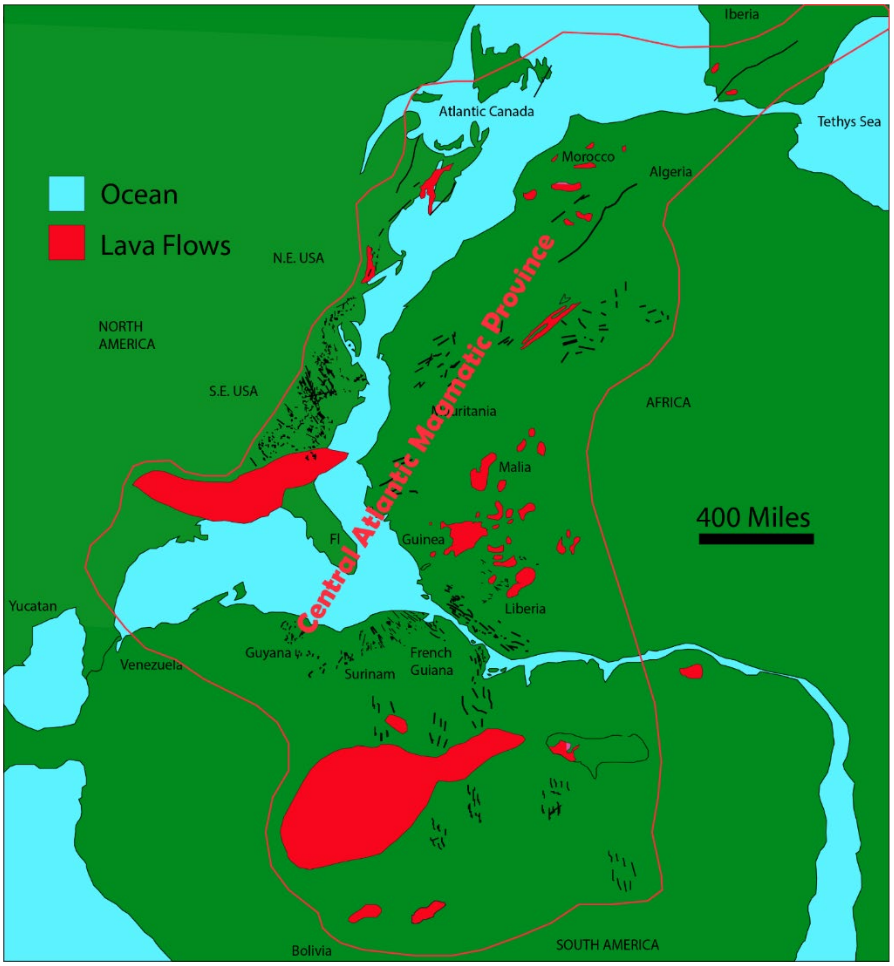


Figure O-15. Paleospastic reconstruction showing the lava flows and dikes (black lines) of the Central Atlantic Magmatic Province (After Hames et al., 2003).

As Pangea continued to rift, separating North America from Europe and Africa, the Atlantic Ocean basin was born (Figure O-14). The basin has continued its growth since 200 Ma, with a mid-ocean rift system active to this day, creating new oceanic crust and enlarging the Atlantic Ocean by about 4 cm per year. The eastern margin of North America reverted back to a passive setting with large volumes of sediment being shed off the continent and deposited along the continental shelf and slope into the ocean basin.

Yet active geologic processes were not finished with northern New England. An anorogenic magmatic event produced some of the most voluminous plutons in the region. This includes the New England Seamounts, a chain of granitic intrusions that form a linear, east-west trend in the Gulf of Maine, the White Mountain Volcanic-Plutonic Suite, a north-northwest belt of plutons that extend across New Hampshire, and the Monteregian Hills, a chain of plutons extending east-west across southern Quebec (Figure O-16).

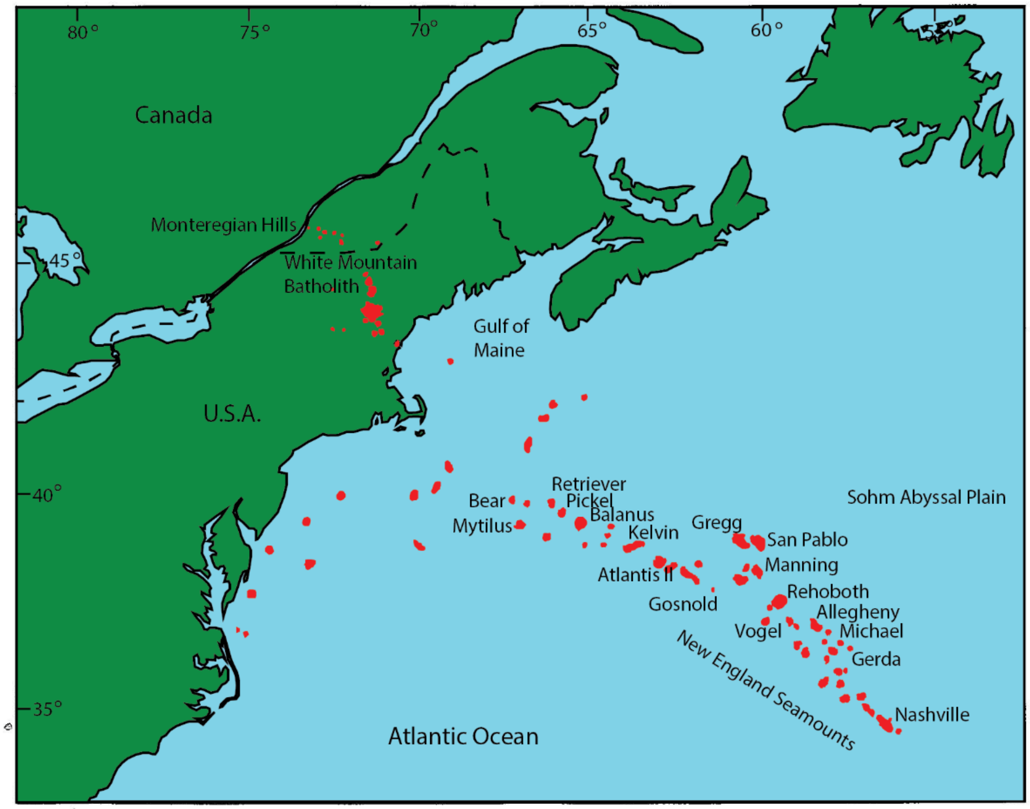


Figure O-16. Simplified geologic map of New England and Maritime Canada showing the location of Jurassic to Cretaceous alkaline intrusions of the Monteregian Hills, the White Mountains, and the New England Seamounts (After McHone, 19996).

The term anorogenic means the magmas were not related to a collisional event like the previously described igneous rocks. Geologists don't agree on the process whereby these granitic rocks formed. Some think a mantle plume impinged along the edge of North America (e.g., Figure I-2). Mantle plumes are thought to be columns of heated rock ascending from the core-mantle boundary. Plumes have lower densities than the surrounding mantle and therefore rise toward the Earth's surface. As the column impinges at the base of the lithosphere, it flattens and partially melts. These basaltic magmas, having high temperatures, cause partial melting of the crustal rocks, generating granitic rocks such as those found in the White Mountains of New Hampshire. Other geologists think the plutons aligned in linear belts resulted from partial melting with melt migration moving up major, preexisting fracture systems. Stops 23-26 are in these anorogenic plutonic rocks.

The net result of all these rifting and orogenic events from ~600 to 100 Ma is displayed in Figure O-17. This simplified geologic map of New England shows the terranes that were accreted to Laurentia since the Taconic Orogeny, ~ 470 million years ago. The Rowe-Hawley and Bronson Hill belts were early additions to the Laurentian margin. Subsequent to the Taconic Orogeny, the microcontinent Ganderia docked during the Salinic Orogeny. These rocks are mostly covered by the younger sediments (now metamorphosed) that were deposited by collapse and extension of the mountain belt led to the development of the Connecticut Valley, Central Maine Basins and the

Merrimack Trough. Sediments were shed into these basins from both Laurentia and from the approaching Gondwanan microcontinents to the east. Avalonia accreted during the Acadian Orogeny; these rocks are best seen in southeastern New England. The Neoacadian Orogeny resulted from the arrival of Meguma, but these rocks are not exposed in New England, lying off shore. To see Meguma rocks, one would have to visit Nova Scotia. For that visit, I highly recommend the field guide by Martha Hild and Sandra Barr listed in the Author's Preface.

The Alleghanian Orogeny, resulting from the joining of this modified Laurenian continent with the Gondwanan continent, metamorphosed many of these accreted terranes that were sandwiched between the two continents. This metamorphism is best seen in southern New England, outside the scope of this book. Acadian metamorphism dominated in northern New England. But the previously accreted Ganderian microcontinent was uplifted during the Alleghanian Orogeny and is exposed through several windows, erosional holes in rocks that were thrust over the Ganderian basement, in New Hampshire as the Massabesic Gneiss Complex (Stop 21). Other Ganderian rocks are exposed through windows in southern New England.

Beyond the scope of this book is the effects of Pleistocene glaciation that strongly influenced the surficial topography of New England.

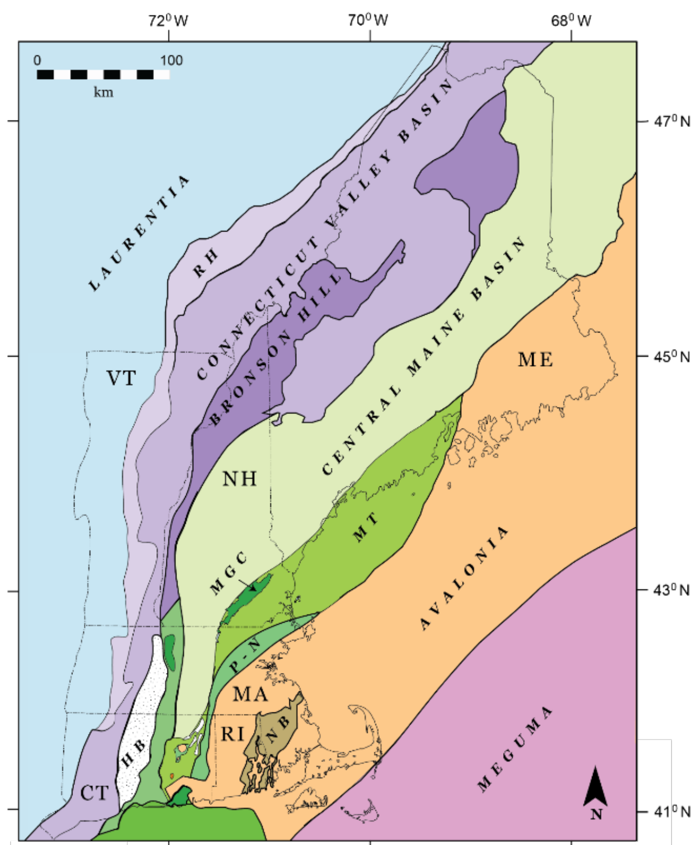


Figure O-17. Simplified geologic map of New England showing the terranes accreted during the Taconic, Salinic, Acadian, Neoacadian, and Alleghanian Orogenies. RH = Rowe-Hawley, MGC = Massabesic Gneiss Complex, MT = Merrimack Trough, P-N = Putnum Nashoba, HB = Hartford Basin, NB = Narragansett Basin

SUMMARY OF VERMONT GEOLOGY

Of the three northern New England states, Vermont has the greatest diversity of geology and rock types. We find rocks formed during the Grenville Orogeny, a series of Late Proterozoic collisional events that created the Rodinian supercontinent, exposed in the Green Mountains (Stop 1). Rifting of Rodinia split Gondwana from Laurentia (Figure O-2), causing mantle upwelling and partial melting. Basaltic magmas intruded as dikes and small intrusions (Figure O-4; Stop 2). Continued rifting formed basins between the newly separated continents. These basins became the depositional sites of immature sediments (Stop 3; Figure O-3)). As the Laurentian margin evolved to the drift state, more mature, quartz-rich sediments were deposited along beaches and offshore (Stop 4). At this time, Laurentia was at tropical latitudes, and the world's oldest carbonate reefs formed (Stop 5). Continued erosion of the Laurentian landforms led to deposition of sediments off the continental margin along the continental platform and slope, extending deeper into the newly formed ocean basin.

About 480 Ma, outboard of Vermont, subduction initiated in the ocean basin and an island arc formed (Figure O-5). Continued subduction consumed the ocean basin, and eventually, the island arc collided with Laurentia. Sediments that were shed off the Laurentian margin were compressed between Laurentia and the island arc. These rocks were deformed, metamorphosed and pushed to the west as thrust sheets over Laurentia (Stop 6; Figure O-7). Vermont hosts one of the best exposed thrust faults in the entire Appalachians at Lone Point on Lake Champlain. Continued compression led to thrust sheets from successively deeper regions to be pushed to the west over the previously emplaced thrust sheets (Stop 7). The arc itself is present in eastern Vermont and western New Hampshire: we'll visit metamorphosed basalts in New Hampshire (Stop 12).

Vermont hosts the only blueschist rocks in northern New England; rocks that were subducted to ~ 30 miles or more. These rocks tumbled back up the subduction zone and were then pushed up over the Laurentian margin like the previously mentioned thrust sheets (Stop 8). Even fragments of the oceanic mantle were broken off and emplaced over the continental margin (Stop 9).

The collision between the island arc and the Laurentian margin thickened, deformed, and metamorphosed the intervening rocks to form the Taconic Mountains. These mountains have long since eroded, but the metamorphic effects of the Taconic Orogeny are well displayed in Vermont. Once the compressive forces ceased, the warm, thickened crust began to collapse, rifting the range and forming the extensional Connecticut Valley trough. This trough became the site of sediment deposition, forming the Waits River Formation (Stop 10) and the Giles Mountain Formation (Stop 11). These formations were metamorphosed in a later collisional event called the Acadian Orogeny as the Avalonian microcontinent collided with the Laurentian margin (Figure O-11).

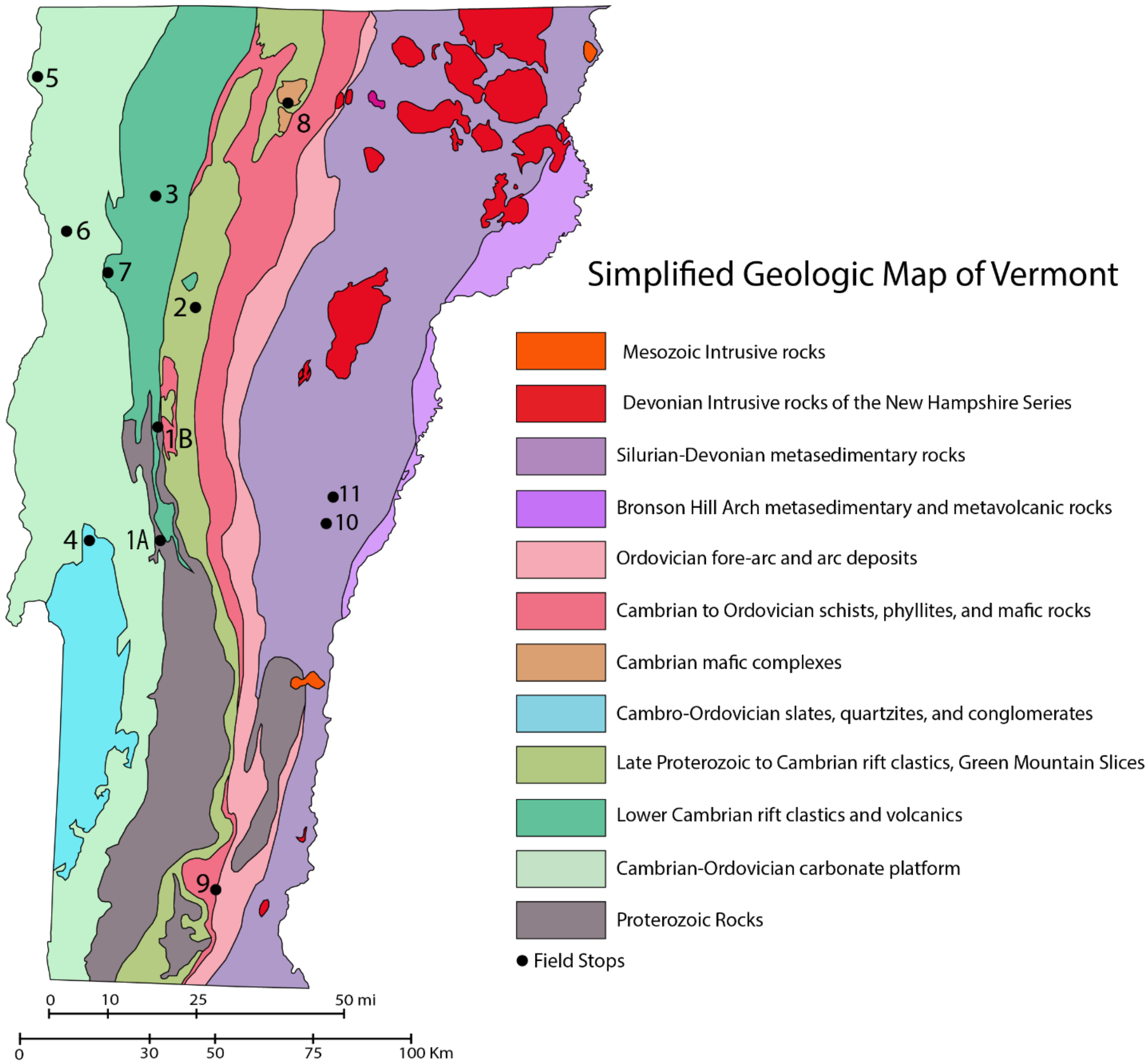


Figure VS-1. Simplified geologic map of Vermont (After Ratcliffe

1a. Mount Holly Complex, Vermont's Basement, Brandon Gap, VT

Most of the surficial rocks of Vermont are sedimentary and metasedimentary rocks of Paleozoic age. But what is underneath these rocks? This stop at the Mount Holly Complex allows observation of what nature is reluctant to reveal – the basement rocks of New England. Geologists use the term basement for rocks that underlie the oldest identifiable rocks in an area, generally crystalline rocks below younger sedimentary deposits. Fortunately, Mother Nature sometimes provides tantalizing glimpses of what she generally hides in northern New England. One of these is the Mount Holly Complex.

These rocks are metamorphosed, deformed granites. They were originally emplaced as plutons into the growing Laurentian crust during the Grenville Orogeny about ~1.3 billion years ago. The Laurentian crust has grown over time as each successive orogeny added additional rocks to its margin (Figure O-1). Much of the geologic story of northern New England details additions to the Laurentian margin from one collisional event after another. The Mount Holly Complex is one such addition from the oldest mountain building event in the northeastern United States.

We owe the Taconic Orogeny for generating thrust faults that brought deep seated rocks to the surface. Subsequent stops in this guide describe classic northern New England thrust sheets of the Taconic Orogeny (Stops 6 and 7) and in a tectonic chronology, this stop would sequentially follow the Hinsdale Thrust of Stop 7. However, because these rocks are among the oldest in Vermont and represent Vermont's basement, I placed this location as our first stop.

Collision of the Iapetus island arc with the Laurentian margin deformed, metamorphosed and thrust rocks that were between the Laurentian crystalline basement and the arc. Most of these rocks were sediments derived from either Laurentia or the arc itself. The collision broke the sedimentary stack into thrust sheets that were transported up and over younger rocks to the west. From west to east in Vermont, these thrusts transported rocks from increasing greater depths. The Mount Holly Complex represents a fragment from the edge of the Laurentian crystalline basement that was transported to the west over younger Paleozoic metasedimentary rocks.

Driving Directions

From Brandon, VT, take RT 73 east (Figure 1-1). About 4 miles east of Brandon, there is a pull off for a picnic table on the right. Good road cuts are on the north side of the road, opposite the picnic area (Figure 1-2). These cuts are downhill from the crest at Brandon Gap which is only 0.2 miles farther east.

Figure 1-1. Map of Route 73 at Brandon Gap showing the two locations described in this stop.

Figure 1-2. Outcrop downhill to the west of Brandon Gap, located across from a picnic area pull off on the south side of the road.

Figure 1.3. Close-up photo of the metamorphosed granitic rocks of the Mount Holly Complex.

On the Outcrop

N43º 50.541’, W 072º 58.445’

The rocks at this outcrop display a well-developed foliation, a planer alignment of biotite, giving the rock a gneissic texture (Figure 1-3). The original, white feldspar crystals have been strained. They no longer have their original rectangular shape but now are tapered at each end as the crystals were stretched. Gneiss formed from metamorphosed sediments is termed paragneiss, whereas metamorphosed, deformed igneous rocks are termed orthogneiss. The overall homogeneous nature of this rock indicates that the parent rock (prior to metamorphism) was homogenous, suggesting that it was originally plutonic and this is an orthogneiss.

Driving Directions

Continue uphill to the east for 0.2 miles to the saddle at Brandon Gap (Figure 1-4). N43º 50.398’, W 072º 58.038’

Figure 1-4 showing the outcrop on the north side of the road at Brandon Gap.

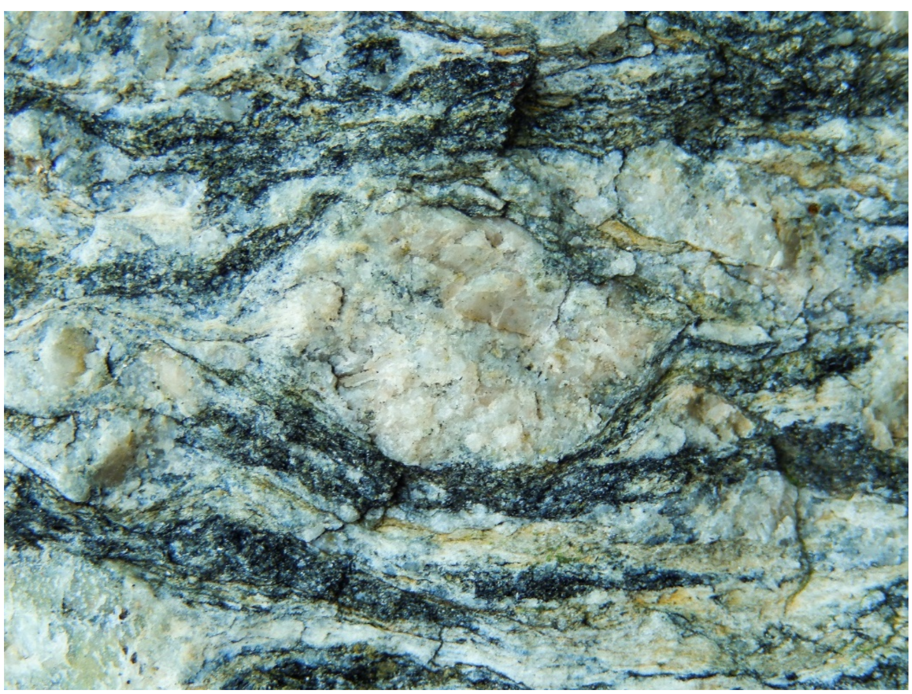


Figure 1-5. Close-up photo of the augen gneiss of the Mount Holly Complex. Tapered feldspar crystal is about ¾ of an inch long.

This road cut shows coarser feldspars than the previous Mount Holly stop. The feldspars are tapered as before, but are much easier to see because of the larger grain size (Figure 1-5). The crystal in this photo was originally like the rectangular Kspar crystals of the Kinsman Granodiorite presented at Stop 17b. These large, tapered crystals are called augen, after the German word for eyes, and the rock is an augen gneiss. The Kspar crystals crystallized from the magma that produced the pluton. Shearing of the rock to the west (to the left of the photo), transposed the rock like a deck of cards as one spreads the deck across a table. The biotite-rich zones represent the boundaries between the cards; i.e., biotite has far less strength than the more rigid feldspar grains and most of the motion occurred along these foliation planes. Shearing reduced the grain size of minerals such as quartz and biotite in these high strain zones, whereas the large Kspar crystals were more resistant to shearing. They tend to break into finer grained aggregates, forming beard-like terminations to the augen.

Significance

How do geologists know that these rocks represent Vermont's basement and are the oldest in the state, especially since they experienced so much metamorphism and deformation? Fortunately, the mineral zircon ($ZrSiO_4$) is very resistant to metamorphism and the age of the original, igneous zircon can be determined. Zircon crystals in the granitic gneiss of the Mount Holly Complex have ages of ~ 1.3 Ga. Some of these grains were abraded to remove the outermost fringes of the grains and were dated at ~ 1.1 Ga. These ages suggest that the granite crystallized earlier in the Grenville Orogeny and was metamorphosed and deformed in a later Grenville event. Therefore, the Mount Holly Complex is part of the Precambrian Grenville orogenic rocks similar to those exposed in the Adirondack Mountains of New York (Figure O-1). The rocks may also have been modified by thrusting during the Taconic Orogeny that broke off this fragment from the basement and brought it to its present location.

1b. Crash Bridge, late Proterozoic Unconformity, Lincoln, VT

Geologists examining road cuts commonly attract the attention of passing motorists. Some honk their horns or yell as they drive by, questioning the sanity of people who examine rocks. Occasionally, karma bites back. A longstanding anecdotal explanation for the name "Crash Bridge" is that a geology class was examining the outcrops along the river and a rude person driving by became distracted and crashed into the bridge.

The previous stop represents a fragment of the Precambrian basement that was thrust upward during the Taconic Orogeny. Likewise, we owe this stop to the Taconic Orogeny as well, but here, the gift of the Taconic was sweet indeed. As mentioned previously, the Rodinian supercontinent amalgamated during the Grenvillian Orogeny, about a billion years ago. In the late Proterozoic, Rodinia began to rift, producing the rift-related basaltic magmas seen at Stop 2 and the immature sediments of the Pinnacle Formation of Stop 3 that were deposited into the rift basin (Figure O-3). Here we see the oldest sediments of the Pinnacle Formation that form its base. These are conglomerates, coarse-grained sedimentary rocks with clasts larger than 1/8 inches in diameter. These rocks have clasts up to 12 inches long. Conglomerates are also seen at the base of the Rangeley Formation (Stop 27). What makes this location interesting is that the conglomerates were deposited directly on the basement rocks, being the first sediments to fill the rifting basin.

Driving Directions

From the traffic lights in Bristol, VT, drive 1.5 miles north on RT 116 to Lincoln Road on the right. Follow Lincoln Road for 4.4 miles to the second bridge beyond Lincoln (Figure 1-6). Park just after the bridge. There is a fisherman's path along the north side of the bridge to descent to the New Haven River.

Figure 1-6. Map showing the location of Crash Bridge near Lincoln, VT

On the Outcrop

Figure 1-7. General view of outcrops under Crash Bridge.

Figure 1-8. Precambrian gneissic basement immediately downstream from the bridge.

Downstream from the bridge, one can see excellent exposures of the Middle Proterozoic basement rocks of Vermont (Figure 1-8). These rocks are part of the Mount Holly Complex as seen at Stop 1a above. They show some differences however, especially since these rocks lack the large, Kspar crystals seen previously. One can see a well-developed foliation formed by the alignment of feldspars and micas. The rather homogeneous appearance of the gneiss suggests that it was originally a granitic body that was metamorphosed during the Grenville Orogeny and perhaps in the Taconic Orogeny as well as the Precambrian blocks were broken from the basement and brought upward with thrust sheets.

Figure 1-9. Photo of the Pinnacle Formation on the upstream side of the bridge.

On the opposite side of the bridge (northern side), one can see the sediments of the Pinnacle Formation (Figure 1-9). This is the same formation seen at Stop 3, but here we seen the basal conglomerates, i.e., the most coarse-grained rocks at the base of the formation. Cobbles of granitic gneiss are immersed as elongated bodies in metagraywacke, an immature sandstone. Previous researchers suggested that the cobbles were deposited during times of very rapid water flow whereas the matrix sediments filled in the spaces between the cobbles during times of reduced flow. The depositional environment may have been an alluvial fan in an arid to semi-arid climate similar to the Basin and Range province of the western United States.

The contact between the basement Precambrian Gneisses and the basal portions of the Pinnacle Formation isn't depositional here. Stress from the Taconic thrusting found a plane of weakness between the two, forming a shear zone at the contact as the Pinnacle Formation rocks moved over the gneisses. Nonetheless, the contact close to here is an unconformity. An unconformity is an erosional surface that separates two rock types of different ages. That erosional surface represents a gap in deposition. This type of unconformity, where sediments are deposited on either metamorphic or igneous rocks is called a nonconformity.

Significance

The basement complexes of northern New England are rarely exposed. The Mt. Holly Complex is one of several basement exposures in Vermont's Green Mountains and represent additions to Laurentian continent during the Grenville Orogeny. But how far east do these Precambrian rocks extend? As will be seen at Stop 21, the Massabesic Gneiss Complex of New Hampshire represents New England's Precambrian basement, but it is Ganderian rather than Laurentian. Ganderia was formed on the opposite side of the Iapetus Ocean and was brought on the oceanic conveyor belt via subduction to collide with Laurentia during the Salinic Orogeny. Somewhere between the Mt. Holly Complex and the Massabesic Gneiss Complex is the suture where rocks on the opposite sides of the ocean are juxtaposed, but because the suture is now covered by younger Paleozoic metasedimentary rocks, the exact location of that suture is speculative.

2. Precambrian Rift Volcanics, Hunger Mountain, Waterbury VT

The supercontinent of Rodinia formed during the Late Proterozoic, about a billion years ago, during the Grenville Orogeny. The previous stop shows granites that were emplaced and metamorphosed during that orogeny. During the Late Proterozoic to Early Cambrian as Rodinia began to rift, the continental crust thinned allowing deep-seated mantle rocks to ascend. Upward migration of the mantle caused partial melting, generating basaltic magmas that flowed into the newly formed faults and fractures that were produced as Rodinian crust was being pulled apart (e.g., Figure O-4). These magmas are now represented as dikes and small intrusions of mafic rocks that are well represented by exposures in Vermont. They have been metamorphosed to amphibolites (metamorphic rocks containing amphibole and plagioclase feldspar of the amphibolite facies, Figure I-6) during the Taconic and Acadian orogenies, but the rocks preserve the chemical characteristics of rift-related magmas that allow recognition of their origin. Subsequently, Rodinia rifted into the smaller continents of Laurentia and Gondwana as the Iapetus Ocean basin formed between them and plate tectonic processes caused the continents to drift apart (Figure O-2).

Driving Directions

Take Exit 10 from I-89 at Waterbury, VT, head north on RT 100 for 2.15 miles. Turn right on Howard Ave for 0.35 miles, then turn left onto Maple Street. Continue for 0.14 miles then turn left on Loomis Hill Road for 3.3 miles (Figure 2-1).

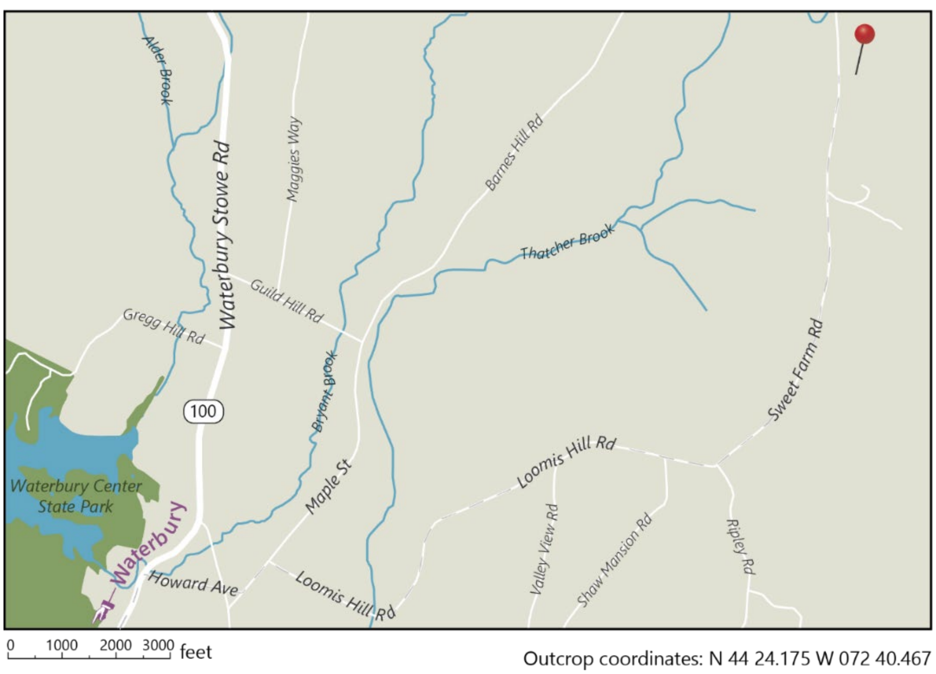


Figure 2-1. Map of the Waterbury, VT area with red pin marking the location of this stop.

Parking Directions

After driving 3.3 miles on Loomis Hill Road, there is a small parking area on the right with a sign for the C.C. Putnam State Forest, Burt Hollow Block, Waterbury Trailhead.

Walking Directions

At the parking lot, follow the trail for 155 feet where a smaller trail branches off to the left. Continue to the top of the quarry (Figure 2-2). N44º 24.175’, W 072º 40.467’

Figure 2-2. View from the top of the quarry of Stop 2.

On the Outcrop

Figure 2-3. Outcrop of amphibolite with plagioclase-rich bands that probably represent segregations of more evolved melt in the original igneous rock.

Late Proterozoic rift-related rocks are represented by dikes and small intrusions such as the one seen at this location. But how do geologists conclude that the amphibolites at this outcrop represent basalts from a rift environment versus any other tectonic setting? Basaltic magmas have compositions that reflect the tectonic setting of origin, for example, subduction zone basalts having subtle chemical differences than basalts produced in rift settings. Upon metamorphism, basalts from both subduction zones and rifts look similar and are commonly metamorphosed to amphibolite facies in orogenic belts, hence their tectonic setting cannot be determined in the field. But chemical analysis of the rocks provides the answer. Even though the original basaltic rock has been metamorphosed to amphibolite facies, the rocks still preserve much of their original chemical composition. Analysis of these rocks and other amphibolites of the same age across Vermont have been compared to the compositions of modern basalts from different tectonic settings across the world and were found to match basalts that have recently or are currently erupting in rift environments. Therefore, geochemistry is a powerful tool in deciphering ancient environments and settings.

Many of the mineralogical and textural features seen here resulted from the Taconic Orogeny; the original mineralogy of the igneous rocks having been completely replaced by metamorphic minerals, the most obvious of which are amphibole and plagioclase. But some relict igneous features include a faint layering present in several locations (Figure 2-3). Some of these are plagioclase-rich, perhaps representing segregations of more evolved magma in the original rock.

Figure 2-4. Quartz-rich veins deposited from metamorphic fluids are also present in the amphibolite.

Very noticeable to the eye are light colored vein-like pods that cut the rock. These are dominated by quartz that represent metamorphic segregations, forming much later than the time of intrusion of the mafic magmas. Quartz veins form when hot, aqueous fluids precipice quartz in fractures.

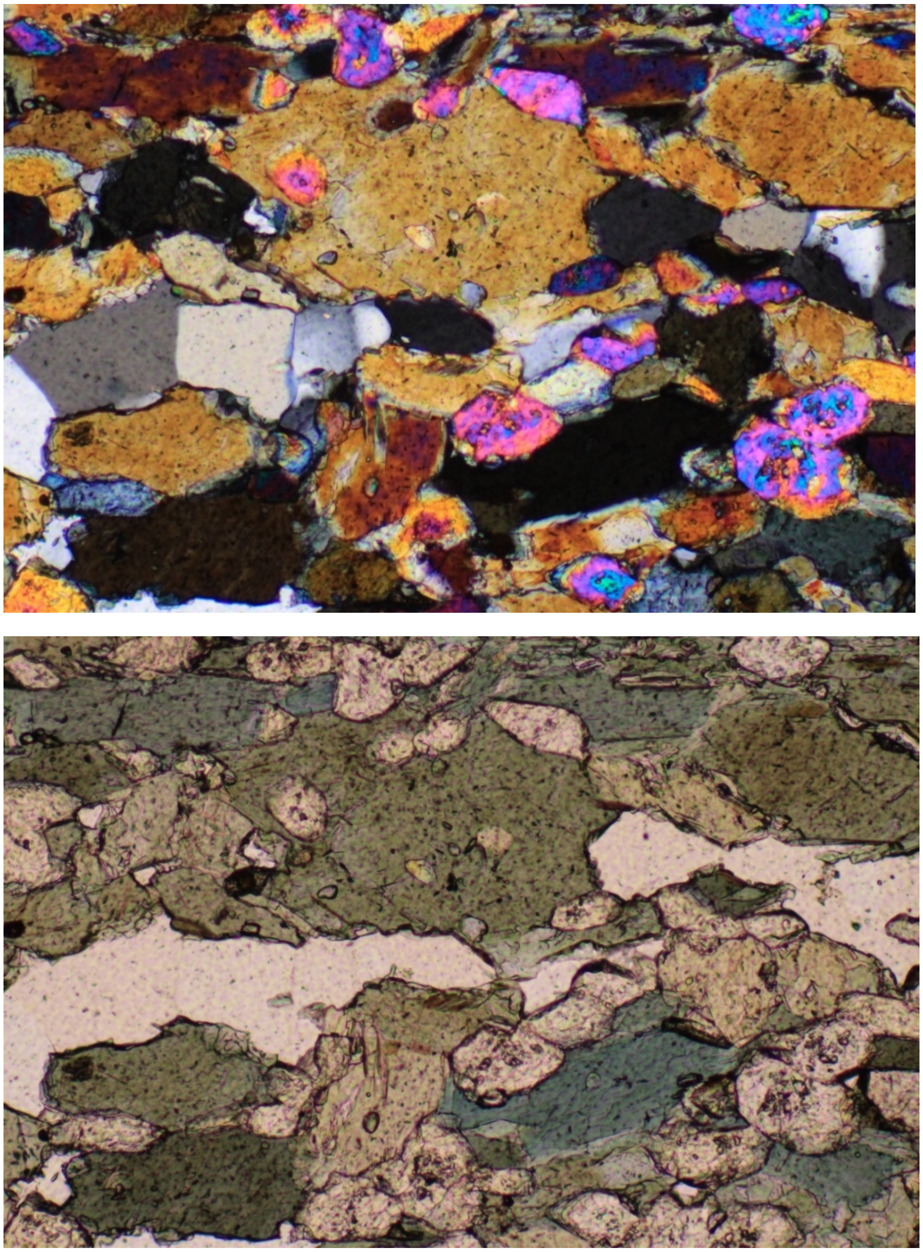


Figure 2-5. Photomicrographs of the amphibolite at Hunger Mountain. The upper photo is a crossed polarized image showing crystals of amphibole (yellow and brown color), epidote (pink to purple color) and a small quartz vein (gray to white color). The lower photo shows the same sample in plane polarized light. The amphibole is green, epidote is a light gray color and quartz is white. None of these minerals were present in the original igneous rock but are the result of metamorphism during the Taconic Orogeny. The presence of both amphibole and plagioclase is characteristic of amphibolite facies metamorphism. Field of view is 5 mm.

3. Late Precambrian Pinnacle Formation, Clastic Rift Basin Sediments, Fairfax Falls Power Station, Fairfax, VT

Late Proterozoic rifting of Pangea caused crustal thinning and upwelling of the mantle and partial melting. Mafic dikes of these mantle-derived magmas are seen at Stop 2. As extension continued, pull apart basins developed and served as depositional sites of sediment derived from adjacent highlands (Figure O-3). These elongated basins are called grabens or rift valleys (Figure I-7D). The sediments of the Pinnacle Formation are excellent representatives of these basin-filling sediments and the exposures at the Fairfax Falls are superb. Unlike sediments described elsewhere in this guide (Stops 27, 28, 29 and 33) that were deposited in deep water in ocean basins, these sediments were deposited in a continental basin prior to full extension and ocean basin formation. As rifting continued, the continental margin matured and the sediments deposited along the newly formed continental margin became progressively richer in quartz. Eventually, the sediments along the margin were dominated by beach deposits dominated by quartz as seen in the Cheshire Formation of Stop 4.

Geologists use the term maturity to describe the texture, composition and the size of grains in clastic rocks. These features vary as a function of the distance of sediment transportation. As sediment is transported, unstable minerals are abraded or chemically break down to leave more stable minerals, such as quartz. Mature sediments consist of grains that are well sorted and well-rounded from weathering and abrasion during transportation and exhibit little compositional variation. Thus mature sediments are more uniform in appearance than immature sediments. Conversely, an immature sediments contain more angular grains, a range in grain sizes, and are compositionally diverse. The sediments at this stop are immature. They represent the early eroded sediments derived from the highlands adjacent to the extending depositional basin. The Pinnacle Formation at Crash Bridge (Stop 1b) has large clasts such that the rock is a conglomerate, having been deposited at the base of the formation. Here at Fairfax Falls, we see sediments higher up in the formation. The grain size is smaller, but the heterogeneous nature of the rocks indicates that they are immature. The reddish-brown color of some of these rocks indicates that iron in the rock is oxidized and the sediments were deposited subaerially, i.e., on land. In contrast, sediments with more greenish colors were deposited subaqueously, i.e., under water where limited oxygen produces reduced iron.

Driving Directions

From the intersection of RT 15 and RT 114, just west of Cambridge VT, drive 5.5 miles on Route 114 to the power station on Lomoille River (Figure 3-1). Drive over the bridge and cross the river and turn left on River Road. At 0.2 miles heading down hill, there is a gravel road that turns sharply to the left. The road may require 4 wheel drive, but space for a couple passenger cars is available where the gravel road joins River Rd. N44º 39.008’, W 072º 59.373’

Figure 3-1. Map of the Fairfax Falls Power Station with red pin marking the location of this stop below the dam.

Walking Directions

Follow the gravel road downhill for a few hundred feet to the river below the dam. Some of the best features are seen on blocks of loose rock that are below the dam rather than on the cliff face itself. Once at the river, walk downstream to examine the many blocks with fresh surfaces. Since the location of an individual block will move depending on stream flow, no specific location is given here. But with minimum effort, you'll find many interesting samples such as the ones illustrated here.

Figure 3-2 showing the general view of the Fairfax Falls Power Station area. Outcrops of interest are in the left side of the photo.

On the Outcrop

Figure 3-3. Immature sediments of the Pinnacle Formation. The white grains are plagioclase, a mineral that does not survive long transportation distances in streams.

The proximity of the source of these sediments is evidenced by immature nature of these rocks, i.e., the sediments have considerable amounts of plagioclase feldspar and lithic fragments that would have been unstable during long transportation distances. Plagioclase crystals up to 1/8 inches are common (Figure 3-3). Transportation in streams tends to abrade and break feldspars, and the relatively well preserved plagioclase indicates that the distance from the source to the depositional basin could not have been very far. Long transportation distances would have broken the feldspars and lithic fragments, and chemical weathering produces clay. These processes tend to concentrate the more resistant quartz grains as the dominant sediment with increasing distance from the source.

Additionally, some boulders at the base of the dam contain large lithic fragments (Figure 3-4). These fragments, called rip-up clasts, consist of finer-grained sediment that were plucked from older deposits and incorporated in younger pulses of coarser-grained sediment. These fragments would have been completely disaggregated if they had long transportation distances.

Figure 3-4. Lithic fragments in the Pinnacle Formation.

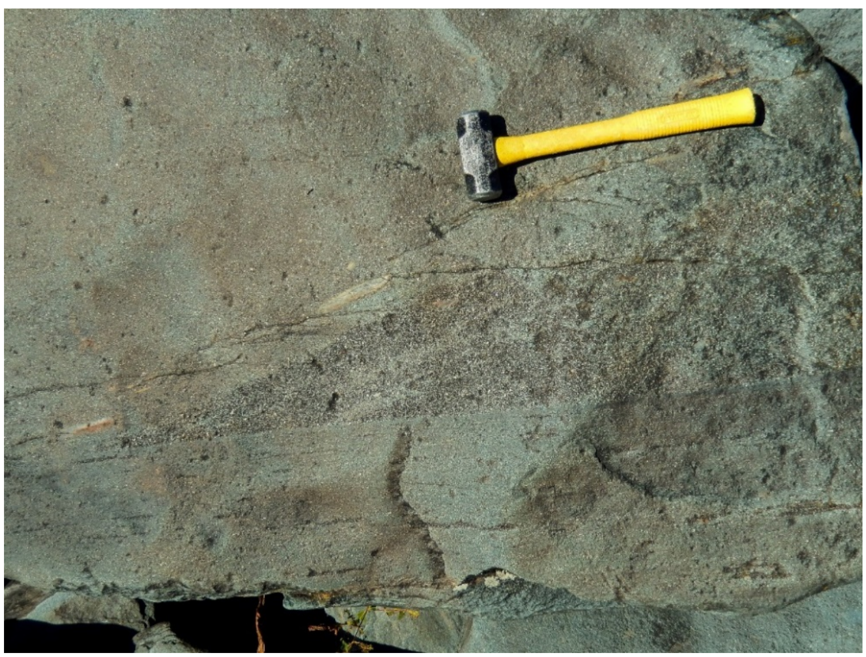


Figure 3-5. Scour features in the Pinnacle Formation.

Other blocks show scour features where streams scoured older sediments. The lowest layer in Figure 3-5 was scoured by the coarser-grained layer in the middle of the photo which in turn, was scoured by another pulse of finer-grained sediment. Note the small, lighter colored, rip-up clasts just above the coarser-grained layer. The clasts are at the bottom of the youngest, upper-most layer in the photo.

4. Lower Cambrian Cheshire Formation at Bristol Falls, Bristol, VT

Vermont is the only landlocked New England state without an ocean shoreline. But this wasn't always the case. Vermont had an extensive ocean shoreline about 550 million years ago. At the end of the Precambrian, continental rifting led to the formation of a passive margin upon which Late Proterozoic to Ordovician sediments were deposited. The term passive margin is a continental margin like the present eastern margin of North America where no active tectonic processes are occurring. The quartz-rich sands of the Cheshire Formation represents the transition from the rift valley sediment such as the Pinnacle Formation of Stop 3 to the development of a carbonate platform deposits as represented by the Dunham Dolomite of Stop 6. The presence of ripple marks in these outcrops of the Cheshire Formation indicate shallow water deposition so they don't strictly represent beach sands, but they represent the quartz-rich sands that were deposited on the tectonically stable shelf. The formation is about 2500 feet thick and records the last stages of transition from the more immature sediments at its base to mature, quartz-rich sediments at the top as seen here. Outcrops of the Cheshire Formation are wonderfully exposed along the New Haven River near Bristol, VT. Vermont may be landlocked now, but during the Cambrian, it had excellent ocean front property.

Directions

From the stop light in Bristol, VT, travel north on RT 116 for 1.5 miles. Just after the second bridge, turn right on Lincoln Road (Figure 4-1). A large gravel pull off is located to the right a couple hundred feet up Lincoln road. Park there and descend to river to the southwest (Figure 4-2).

Figure 4-1. Map of Bristol Falls at Bristol, VT. Red pin shows the locations of the outcrops of interest.

Figure 4-2. General view of the rocks at Bristol Falls. N 44º 07.723’ W 073º 03.010’

On the Outcrop

Figure 4-3. Well displayed bedding of the Cheshire Formation at Bristol Falls.

A very noticeable feature of these outcrops is the very well-defined sedimentary layering. On average, the bedding is 4-6 inches thick. Note that in Figure 4-3, the rock is predominantly composed of quartz with thinner layers that were more clay-rich – these are the thin, darker layers seen in the figure. The grain size of the sediment was not coarse and is dominated by quartz without much feldspar, indicating that the sediments were transported a fairly long distance. During transportation, larger grains of feldspar were abraded and broken, altering to clays during weathering. These sediments are mature compared to the immature Pinnacle Formation of Stop 3.

Figure 4-4. Ripple marks at the top of a bed in the Cheshire Formation.

Looking down at the tops of the beds shows ripple-like features (Figure 4-4). Ripple marks are formed from flowing water and the asymmetry of ripples indicates water flow direction. Water carries sediment up a gradual slope to the crest of the ripple and then the grains tumble down the steeper backside slope. Figure 4-4 shows ripples with shallow slopes to the upper left and steeper slopes to the lower right, indicating a current direction from the upper left to lower right.

The sediments that formed these rocks were not dissimilar to those of present day's northern New England's sandy beaches at Hampton, Rye, Ogunquit, etc. What was once a passive continental margin in Vermont about 550 million years ago has reverted back again to a passive margin today. Eventually, a subduction zone system may initiate in the Atlantic Ocean basin and New England could again be the site of volcanos and mountain building, collisional events. And who knows, perhaps in another 550 million years, new beach front property could form in northern New England.

5. Isle La Motte, Chazy Reef, Goodsell Ridge Preserve, La Motte, VT

The first European visitor to Isle La Motte was Samuel de Champlain in 1609. It was the site of Vermont's first European settlement when Pierre La Motte built a military outpost on the island in 1666. General Richard Montgomery's troops camped on the island during their failed invasion of Canada during the Revolutionary War. Likewise, General John Burgoyne's invasion army camped there prior to the British defeat at the Battle of Saratoga in 1777. But there were far older inhabitants of this location. Isle La Motte hosts the world's oldest reef complex and contains abundant Lower Cambrian fossils.

As Rodinia rifted during the Late Proterozoic, the newly formed Laurentian margin evolved from continental pull apart basins that were filled by sediments from the highlands immediately adjacent to the basins (i.e., the Pinnacle Formation of Stops 2 & 3). With time, the sediments became more quartz-rich (Cheshire Formation of Stop 4) as the edge of Laurentia evolved to a passive margin. As the margin continued to evolve, carbonate platforms developed about 480 million years ago. These platforms were similar to those located in the Bahamas and off the southern Florida coast. Laurentia was located south of the equator at this time, and warm, tropical seas gave rise to what is now the world's oldest preserved coral reef. The reef system originally was extensive along the Laurentian margin, but now only small portions remain, the best being at Ilse La Motte.

These Cambrian reefs, named the Chazy Group, are well displayed at Goodsell Ridge Preserve. At the preserve, you'll see not only the coral reefs themselves, but also many fossils representing organisms that dwelled among the coral reefs. The sedimentary rocks on Ilse La Motte have been gently tilted, showing a cross section through time and a glimpse of the evolution of life during the Cambrian Period. Farthest south on the island, the oldest rocks preserve bryozoan fossils. Stratigraphically higher, younger corals appear, followed by stromatopoids, a type of sponge. At the Goodsell Ridge Preserve itself, the rocks represent an explosion of biodiversity, preserving evidence of a complex ecosystem with bryozoan, receptaculitid algae, nautiloids, sponges, corals.

An overview of the significance of the Isle La Motte reefs is found at the Ilse La Motte Preservation Trust web page:
http://www.ilmpt.org/ilmpt/Welcome_to_Isle_La_Motte_Preservation_Trust.html.

Driving Directions

From Burlington, travel north on I-89 to Exit 17 and drive west on RT 2. Cross causeway to South Hero and continue on RT 2 past the town of Grand Isle. Proceed through the town of North Hero, following RT 2. Just after the causeway leading from North Hero Island, turn left on RT 129. Follow Route 129 to causeway leading to Isle La Motte. Cross the causeway to Isle La Motte and drive 4.0 miles through the village of Isle La Motte to Quarry Road on the left. Turn left and drive 0.2 miles to the Goodsell Ridge Preserve on the left (Figure 5-1).

Figure 5-1. Map of the Goodsell Ridge Preserve at Ilse la Motte, VT.

Where to Park

There is ample parking at the Preserve near the Visitor's Center.

Walking Directions

From the Visitor's Center, follow the trail to the various Discovery Areas described below.

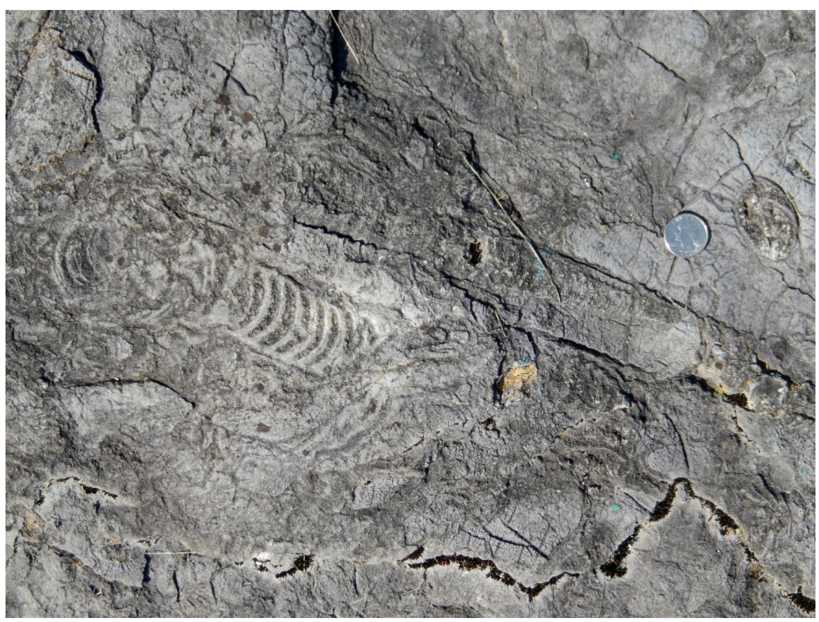


Figure 5-2. Cephalopod fossils located just west of the Visitor's Center (N 44º 51.163' W 073º 20.418').

About 20 feet west of the Visitor's Center, a ground level outcrop of limestone displays very nice cephalopod fossils (Figure 5-2). These fossil cephalopods represent animals that dwelt in chambered shells and are related to modern-day squids and octopi. This fossil preserves the shell of the organism; the head of the cephalopod would have extended out the end of the tube-like shell like modern squids. Note the conical shape of this early cephalopod. Cephalopod shells later evolved into a curved shape, best represented by ammonites that became extinct at the end of the Cretaceous and by modern nautilus species. This rather benign looking fossil was actually near the top of the food chain in its day, using tentacles to capture their prey.

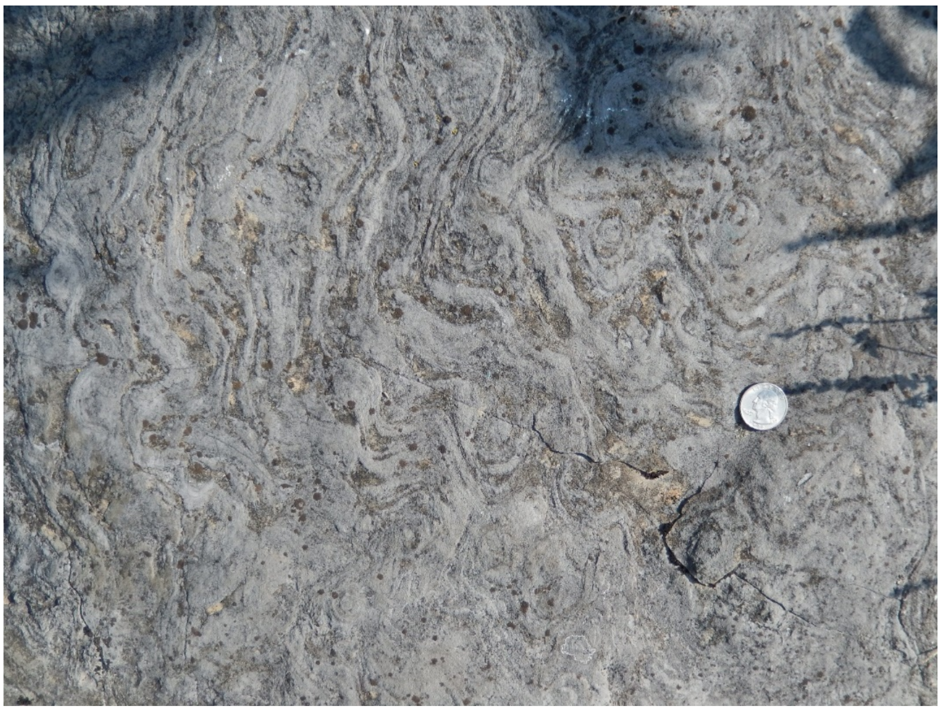


Figure 5-3. Stromatoporoid at Discovery Area 4 (N 44º 51.197' W 073º 20.445').

Follow the White Trail to Discovery Area 4. Stromatoporoid colonies are displayed here (Figure 5-3). Previously thought to be corals, stromatoporoids are now recognized as calcareous sponges. Together with corals, they became the main reef builders during the Middle Paleozoic, living in shallow water, shelf settings. The circles that look like bulls eyes are mamelons, projections that allowed the stromatoporoid colony to have greater contact with wave currents and to more easily capture food. They created currents to pump water in and out of their bodies where they filtered very small food particles.

Figure 5-4. Gastropods at Discovery Areas 5 & 6 (N 44º 51.231’ W 073º 20.443’).

Continue on the trail to Discovery Areas 5 and 6. Here you’ll see fossils of gastropods (Figure 5-4). If you think these look like snail shells, you are correct, snails are a modern type of gastropod. But unlike snails, these creatures obviously could handle salt, in fact, the first gastropods were exclusively marine animals. Note the large size – the shells were up to 4 inches in diameter.

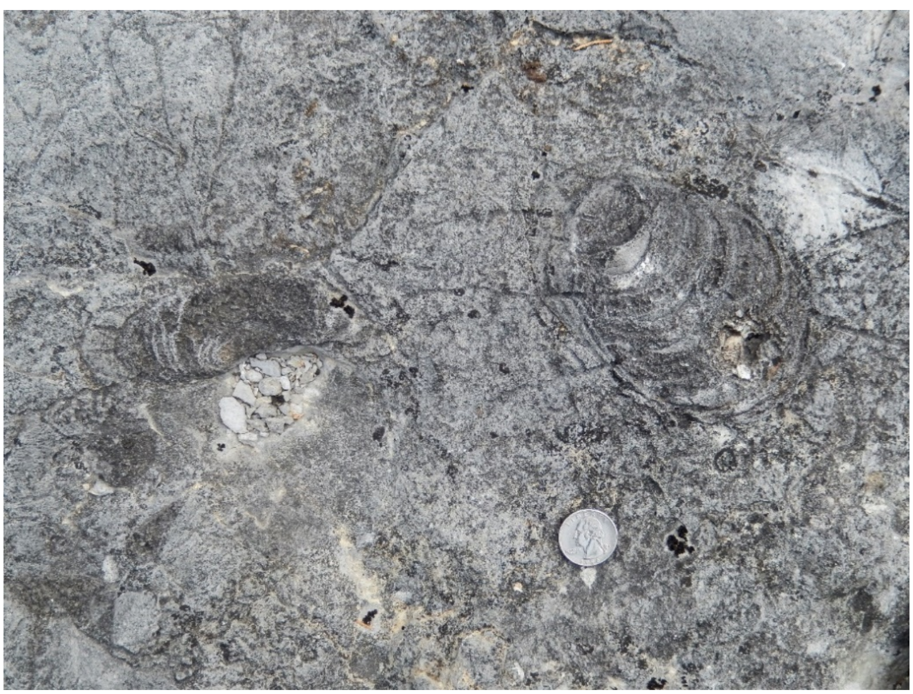


Figure 5-5. Tabulate corals (N 44º 51.266’ W 073º 20.447’).

Here you can also observe the fossils of tabulate corals (Figure 5-5). This type of coral lived as colonies in shallow water and were relatively rare until the Ordovician period. They began to decline in abundance in the middle of the Silurian period until becoming extinct at the mass extinction event at the end of the Permian about 252 million years ago.

The entire coral as seen here is called the corallum which consists of colonies of many individual tabular chambers called corallites. At the base of individual corallite chambers are thin plates called tabulae after which the coral group is named. The actual coral animal is a sac-like animal called a polyp that lived in the open, cup-like portion of the chamber, extending venom-bearing tentacles for capturing food to draw to its mouth at the base of the tentacles. These small animals secreted calcium carbonate to build a hard skeleton that forms the coral mound. Over time, a series of chambers built upward with abandoned old living chambers below.

Other Locations: Fisk Quarry Preserve

Driving Directions

From the Goodsell Ridge Preserve, return to Quarry Road and turn left (Figure 5.1). Drive for 1.5 miles to the Fisk Quarry Preserve sign. Park at N 44º 50.695' W 073º 21.705'.

This and 5 other quarries on Ilse La Motte were worked in the 19th century. Limestone from these quarries was used to construct the Brookline Bridge, Radio City Music Hall and the National Gallery of Art.

In the walls of this quarry, you'll see excellent cross sections through stromatoporoid sponges. Along the flat rock surfaces to the south of the quarry pond, many good cephalopod fossils are found. At the time of my visit, piles of small rocks form rings around good fossil locations.

6. Champlain Thrust, Thrust Fault Extraordinaire, Lone Rock Point, Burlington, VT

The Champlain thrust fault that is spectacularly exposed along the shore of Lake Champlain at Lone Rock Point could very well be the most famous outcrop in New England and one of the best exposed thrust faults in North America. We can see the older, Lower Cambrian Dunham Formation that was trust over the younger, Middle Ordovician Iberville Formation during the Taconic Orogeny. The features exposed here make this outcrop a structural geologist's delight.

During the Taconic Orogeny, an island arc that was located in the Iapetus Ocean collided with the Laurentian margin (Figure O-7). Consider this collision to be like a large truck smashing in a line of cars waiting at a stop light. Just as the cars would be smashed and pushed forward, so were the rocks lying between Laurentia and the colliding island arc, albeit the movement was at geologic rates. This collision produced a series of thrust sheets that originally were at deeper crustal levels and ended up being pushed over younger rocks to the west (Figure I-7E). The Champlain thrust is the oldest of several thrust sheets in Vermont that were activated in the Taconic Orogeny. Here, the older, Cambrian Dunham Formation may have moved up to 50 miles from its initial position, having been trust to the west where it now overlies younger, Ordovician rocks of the Iberville Formation. Another thrust called the Hineburg thrust (Stop 7), containing rocks from a deeper location farther to the east, was pushed over the rocks of the Champlain thrust. Finally, a very deep-seated thrust brought up portions of the crystalline Precambrian basement and transported them to the west over younger rocks (Stop 1).

Driving Directions

From the center of Burlington, VT, drive several miles north along North Avenue, also marked as Vermont State Road 127. At the traffic light at Institute Road, turn left (west) and drive past the Burlington High School (Figure 6-1). Just after the high school, turn right (north) on Rock Point Road that has a gate made of two brick pillars and signs for "Rock Point Episcopal Center". Continue to the Episcopal Diocesan Center.

Figure 6-1. Map of Lone Rock Point, Burlington, VT

Where to Park

Park at the lot just past the baseball field on the right. During normal weekly office hours, stop at the Administration Building to obtain permission to proceed to the lake shore. The Center is very obliging to geologists who desire to visit the outcrops and obtaining permission is never a problem.

Where to Walk

From the parking lot, continue down the road to the north. At 500 feet, you'll cross a bridge over an old railroad bed that is now a bike trail. Take the left fork after the bridge. Pass the Bishop's house on right at 0.13 miles and follow the gravel road. At 0.38 miles, pass the farmhouse on the right. Follow the foot trail to the left. At 0.47 miles to the west of the farmhouse, you'll see a sign with an overview of Lone Rock Point. Take the left branch of the trail at this point. Continue to 0.68 miles where the trail descends down to the lake. Continue to the north at 0.70 miles to excellent exposures of the Champlain Thrust (Figure 6-2; N 44° 29.449'; W 073° 14.930').

Figure 6-2 showing the Champlain Thrust with the upper Dunham Formation and the darker Iberville shale at the base.

On the Outcrop

First notice the long planar surface separating the darker colored, intensely contorted Iberville Formation at the lower portion of the outcrop from the more massive, less deformed Dunham Formation that forms the cliffs above. This surface is the Champlain Thrust fault. The darker colored Iberville Formation at the lower portions of the outcrop consists of shales that were easily deformed as the Dunham Formation slid over the Iberville. At greater distances from the fault, the Iberville shales shows bedding and folds; these are increasingly disrupted with proximity to the thrust fault to where most of the original bedding and folds have been obliterated and the rock now showing many broken and transposed calcite bands (Figure 6-3).

The Dunham Formation consists of a carbonate rock known as dolostone, and being more rigid than the Iberville Formation rocks, it shows far less deformation. The bottom side of the Dunham Formation shows what may appear to be upside-down wave-like features (Figure 6-4). These are called fault mullions and were formed as the Dunham dolostone moved over the underlying rocks and was in essence, scratched in the process. Note the orientation of the “waves” and “troughs”. They align in a N60°W direction, showing the transportation direction of the Dunham Formation during thrusting.

Figure 6-3. The Iberville shale near the thrust plane is highly contorted from the shearing effects of the Dunham Formation overriding the less competent shale.

Figure 6-4. The Champlain Thrust fault and mullions at the base of the Dunham Formation.

7. Hinesburg Thrust, Mechanicsville, VT

Rift-related clastic sedimentary rocks, such as the Pinnacle Formation of Stop 3, were deposited during the earliest rifting of the supercontinent Rodinia (Figure O-3). As the continental margin evolved with the opening of the Iapetus Ocean, more mature sediments were deposited on the continental shelf as the Cheshire Formation (Stop 4), and the carbonates of the Chazy Group (Stop 5) and the younger, overlying formations of the stable platform. Throughout this time, sediments were carried out to sea and were deposited along the continental slopes into deeper ocean waters. During the Taconic Orogeny, several thrust sheets of these marine sediments were pushed to the west during the Taconic Orogeny as the island arc collided with the Laurentian margin. From west to east, the thrust sheets originated from progressively deeper in the crust and consist of rocks that were formed from sediments farther to the east (Figure O-7). The thrust at Champlain thrust at Lone Rock Point (Stop 6) has rocks that were never very deep; these lack any metamorphic effects from higher pressure. The rocks of the Hinesburg thrust originated from deeper levels than the Champlain thrust, and show moderate effects of metamorphism. At this location, we view the Lower Cambrian Cheshire Formation that was thrust over the younger, Ordovician Bascom Formation. These rocks preserve several sense-of-shear indicators, i.e., features that indicate the direction of movement during thrusting. These are especially well preserved along the shear zone between the two formations. Here we examine these sense-of-shear indicators.

Driving Directions

From I-89, take Exit 12 and follow Route 2A for 4.9 miles to the junction with VT Route 116. Turn south on 116. At 7.7 miles, turn left on Mechanicsville Road, heading east toward Richmond. At 8.6 miles, proceed through a 4 way stop intersection, and follow Pond Road. At 9.3 miles, turn left on Place Road (Figure 7-1). A small dirt road turns right off Place Road at 0.1 miles. Park at the first telephone pole along the dirt road.

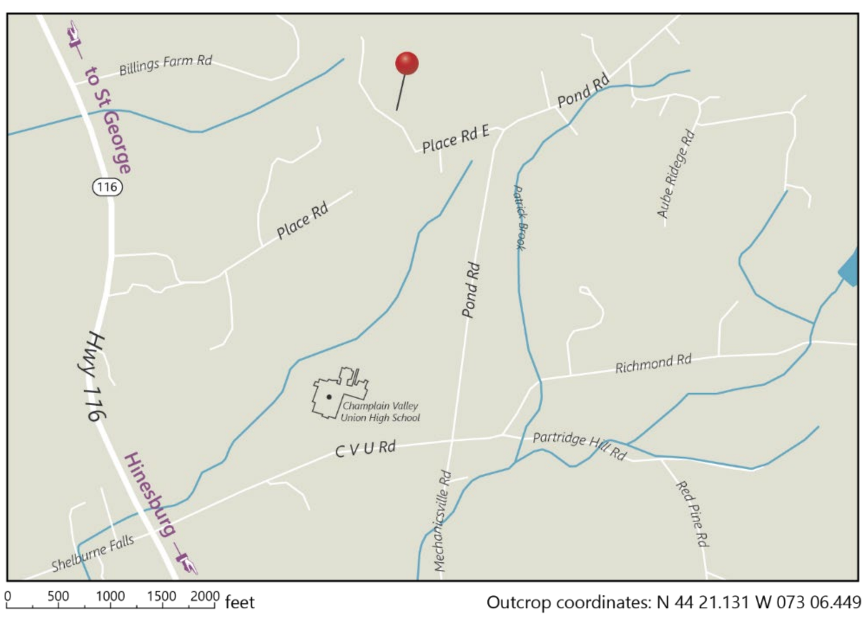


Figure 7-1. Map of the outcrops of the Hinesburg Thrust, Mechanicsville, VT.

Walking Directions

Walk farther along the road for 170 feet from 1st telephone pole and enter woods on the right. About 80 feet from the road at N44º 21.131' W073º 06.449', you'll see a plaque designating this outcrop as a Rolfe Stanley Teaching Outcrop.

On the Outcrop

Figure 7-2. The thrust plane forms the contact between the older, Cheshire Formation (above) and the younger, Bascom Formation (lighter colored rocks below). GPS devise for scale on the Bascom Formation.

Unlike the Champlain thrust of Stop 6 that was never very deep in the crust, the Hinesburg thrust (Figure 7-2) occurred at deeper crustal depths and the rocks have been moderately metamorphosed to chlorite grade of the lower greenschist facies (Figure I-6). These greater depths also meant that the rocks that define the shear plane between the two rock formations was subjected to greater pressure and were hotter than those of the Champlain thrust. As a result, the rocks in the shear zone were more ductile and preserve evidence of flow as the overlying rocks slid over them to the west. Figure 7-3A shows small folds that are stacked upwardly to the left of the photo. Figure 7-3B shows two folds with the same orientation, the upper one folded to the upper left, the lower folded to the lower right. This configuration means that the upper portion of the rock moved to the upper left compared to the lower portion of the rock. The direction of these folds preserve evidence that as the overlying Cheshire Formation moved over the underlying Bascom Formation, the interlaying shear zone was smeared between them, indicating a westward motion of the thrust.

Figures 7-3A and 7-3B. The shearing effects of the overriding Cheshire Formation folded and deformed the layers of the Bascom Formation.

8. Blueschists: Rocks Deeply Subducted in the Ordovician Subduction Zone, Tillotson Mountain, Lowell, VT

Prior to the Taconic Orogeny, eastward subduction consumed the oceanic basin between Laurentia the off-shore island arc. But what happened to subducted oceanic crust? The minerals in the oceanic plate experience a continual change in pressure and temperature as the plate descended. When the plate reached depths of 15 to 30 miles, the basalts were metamorphosed at high pressures and low temperature (400 to 750 degrees Fahrenheit) to blueschists (Figure 1-6). The name refers to the presence of a bluish-colored amphibole called glaucophane whose abundance imparts the same color to the rock.

Most of the subducted oceanic crust continued to descent deeper into the Earth's mantle, but sometimes, fragments of the subducted material tumble back up the subduction. Such is the case with the blueschists found at Tillotson Mountain.

Blueschists are rare in New England; this rarity is why this stop is included in this guide. One criterion in picking field stops for this guide was ease of access, but this policy was abandoned for this location because a hike of several miles is necessary to visit the blueschists. Some visitors may question if the hike is worthwhile to see such fine grained rocks. But don't let the fine-grain size of the blueschists at Tillotson Mountain fool you; the significance of these rocks is considerable, representing a unique geological occurrence in all of New England.

Driving Directions

Take Vermont Route 100 past Eden to Eden Mills. At Eden Mills just past the general store, turn left (north) on North Road. Drive 5.0 miles to Tillotson Road on the left. Follow Tillotson Road for 0.6 miles to parking lot on the left.

Figure 8-1. Map of the Tillotson Mountain area showing the parking location and the outcrop location for this stop.

Walking Directions

One can either follow the trail or the abandoned road to where they cross at about 0.1 miles, then continue on the trail. At N44° 47.615’ W072° 31.761’, the trail junctions. Take the right branch and continue to the Tillotson Camp (Figure 8-2).

Figure 8-2. Hut at outcrops of interest at Tillotson Mountain.

On the Outcrop

Figure 8-3 shows a blueschist outcrop adjacent to the hut on Tillotson Mountain. The rocks are fine-grained, and the visitor may be disappointed that one cannot discern the individual minerals in the rock without magnification. For this reason, the photomicrographs of Figure 8-4 are included. These show fine-grained glaucophane, the Na-rich amphibole indicative of high pressure metamorphism, and garnet. In spite of the fine-grain size, the overall bluish color of the rocks is obvious.

Figure 8-3. Bluish colored outcrop at Tillotson Mountain.

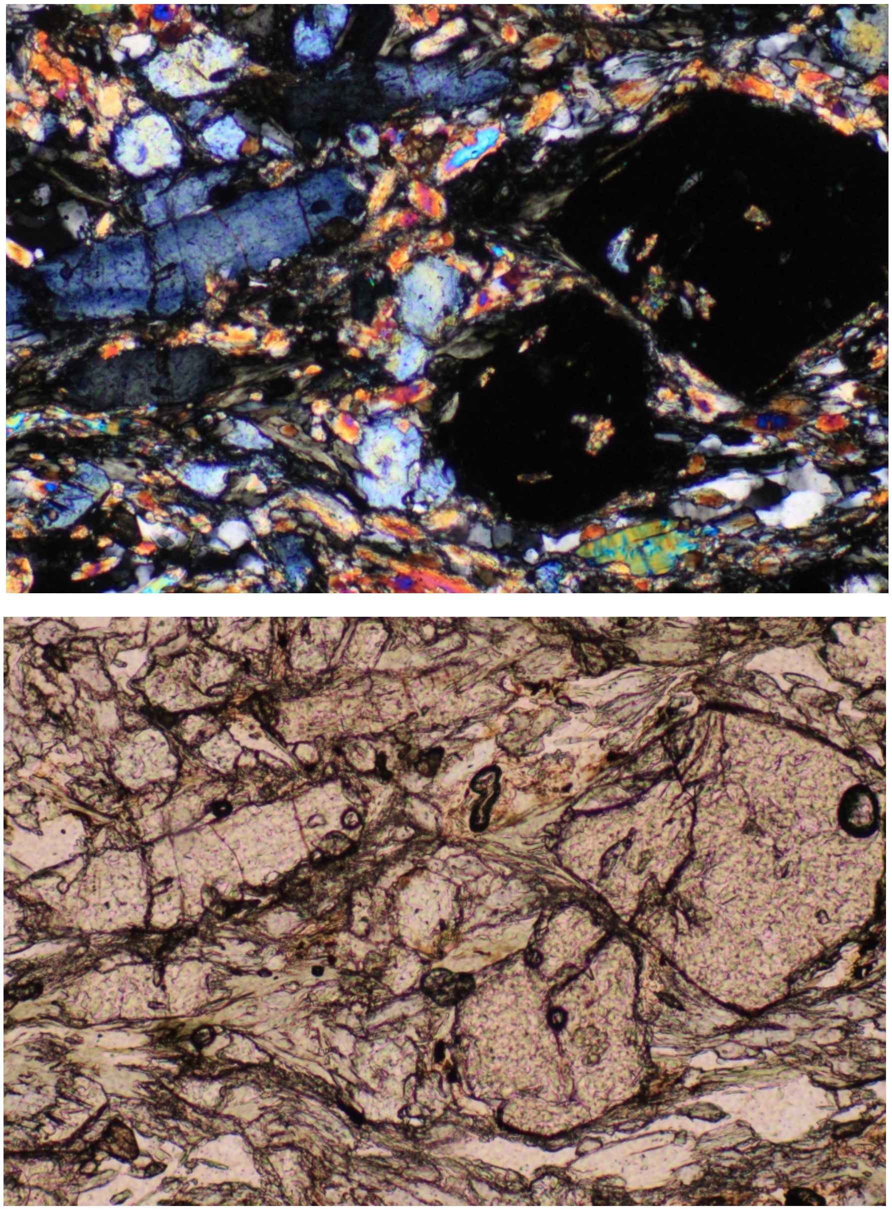


Figure 8-4. Cross polarized light (top photo) and plane polarized light (lower photo) photomicrographs of Tillotson Mountain blueschist. In the upper photo, the bluish mineral in glaucophane, the black mineral is garnet. Field of view is 2 mm.

Blueschists form at depths of 15 to 30 miles below the Earth' surface. Cold oceanic crust, consisting mainly of basaltic rocks, descends rapidly to those depths and because rocks are poor conductors of heat, the rocks remain relatively cold, reaching temperatures of 400 to 750 degrees Fahrenheit. Rocks of the blueschist facies therefore represent a low temperature, high pressure facies. Well exposed blueschists are present in the Franciscan Formation of California, and in Geece, Turkey, Japan, New Zealand and New Caledonia. Tillotson Mountain is unique in New England, being the only blueschists in New England.

Recent discovery of coesite, a high pressure form of quartz, in these rocks indicates that the Tillotson blueschists were subducted considerably deeper than initially thought. Coesite becomes stable at depths of ~ 50 miles, rendering ultrahigh pressure metamorphic conditions.

The preservation of the blueschist facies mineral assemblage requires two specific geologic conditions. First, the rocks cannot descent deeper than the blueschist field shown in Figure I-6. Continued subduction of the oceanic crust will pass the rocks through the blueschist facies into the eclogite facies indicative of higher temperatures and high pressures. Secondly, for blueschist facies rocks to return to the Earth's surface, the rock must be transported upward swiftly. The rocks cannot remain at depths of 15 to 30 miles for very long because eventually, the rocks will heat up sufficiently to transform to either the eclogite or greenschist facies and the mineral assemblage will change to reflect these new pressure – temperature conditions.

Returning relatively high density blueschist rocks to the Earth's surface is problematic. They are thought to migrate upwardly by flow and/or faulting in upper parts of subducted crust, especially if they are mixed with low-density continental crustal rocks such as marble, metasedimentary schists and serpentenites (Stop 9). Blueschists are commonly found in melanges, large-scale breccias from subduction zones. While the blueschists may have high densities, the mixture of rock types in the mélange reduces the overall density, providing buoyancy for the entire package and allowing transportation towards the Earth's surface.

The reader may be disappointed with the fine grain size of the minerals in the Tillotson blueschist. Grain size in blueschists is rarely coarse because mineral growth is hindered by the lack of time at depth. The rapid rate of the rock's metamorphic trajectory, combined with the low temperatures of metamorphism, prohibit large crystal formation. Small grain size notwithstanding, the blueschists of Tillotson Mountain are a New England gem and well worth the visit.

9. Ultramafic Rocks: Mantle Rocks at the Surface, East Dover, VT

Ultramafic rocks are derived from the Earth's mantle from depths below the continental and oceanic crust. The majority of the mantle is composed of peridotite, a rock that mainly contains olivine with lesser amounts of pyroxene. With large degrees of partial melting, basaltic magmas were extracted from mantle peridotite to produce the Taconic Island Arc and other basalts, leaving a residual rock even richer in olivine. These residual mantle rocks that are very olivine rich and are called dunites. Metamorphosed dunite is abundant at East Dover where the original rocks contained over 90% olivine. Subsequently, the olivine has been altered, but the question is: What is a mantle rock doing at the Earth's surface in the middle of Vermont?

As the Taconic Orogeny thrust the island arc over the edge of the Laurentian margin, numerous thrust sheets were pushed to the west as described at Stops 1, 6, and 7, each originating deeper in the crust than the proceeding thrusts, even to the degree that continental crystalline basement was thrust up as the Mount Holly Complex of Stop 1. But mountain building events are not delicate; a portion of the oceanic crust was broken and pushed up by the colliding island arc as well. In this case, a broken portion of oceanic crust in the Iapetus Ocean between Laurentia and the island arc was pushed over the continental margin. The break was deep enough that mantle rocks that were below the oceanic crust were fragmented and now reside at the surface for us to examine. The outcrops at East Dover are part of a continuous belt of ultramafic rocks that extend from Newfoundland to Alabama that were emplaced during the Taconic Orogeny.

The best location to see ultramafic rocks in Vermont, and perhaps in all of New England, is at the Belviedere Mine at Lowell. Unfortunately, the mine is closed and not open to the public. But if it ever opens, Belviedere is definitely the place to go. The mine was active for most of the 20th century, being a major supplier of chrysotile asbestos. The health concerns with asbestos led to the closure of the mine in 1993. A good review of the geology of Belviedere Mountain is found on the Vermont Geological Survey web page: http://www.anr.state.vt.us/dec/geo/bmtn.htm

Driving Directions

From East Dover, VT, drive east on Dover Road for 0.2 miles (Figure 9-1). It's easiest to park at the top of the hill on the south side of the road.

Walking Directions

From the pull off, walk down hill to road cuts on the north side of the road about half way down the hill (Figure 9-2).

Figure 9-1. Map of the East Dover area with red ovals showing the locations of the outcrops.

1) On the Outcrop

The ultramafic rocks at East Dover are highly altered. Olivine is a mineral that is not in equilibrium at the Earth's surface, readily transforming under metamorphic conditions to a hydrous mineral called serpentine. Ultramafic rocks that are dominantly serpentine are called serpentinites. As these initially, hot mantle rocks were thrust over the continental margin, they came in contact with water, transforming olivine to serpentine. The amount of serpentinization of the rocks at East Dover varies, but most rocks contain greater than 50% serpentine; only minor amounts of original olivine left. The outcrops are massive with variable degree of weathering on the rock surfaces.

Figure 9-2. General view of the outcrops of Stop 9-1. N42º 57.011', W 072º 45.695'

Driving Directions

From these road cuts, drive ~ 0.3 miles west to the fork at East Dover (Figure 7-1). Take the left fork for 0.1 miles to the East Dover Fire Department and park.

Walking Directions

Cross the road exactly opposite the Fire Station and proceed about 30 yards down to the stream.

Figure 9-3. General view of the ultramafic rocks facing south from the East Dover Fire Station. N42º 57.011’, W 072º 45.695’

Figure 9-4. River outcrop of ultramafic rocks.

2) On the Outcrop

These ultramafic rocks are similar to those described above, but the stream has polished them allowing better examination of the mineralogy (Figure 9-5). Most of the rock consists of serpentine, the brownish colored mineral. But if one looks carefully, equant, darker minerals can be seen immersed in a sea of serpentine (Figure 9-6). These are olivine from the original rock that were only partially converted to serpentine.

Figure 9-5. Close-up photo of ultramafic rock at East Dover.

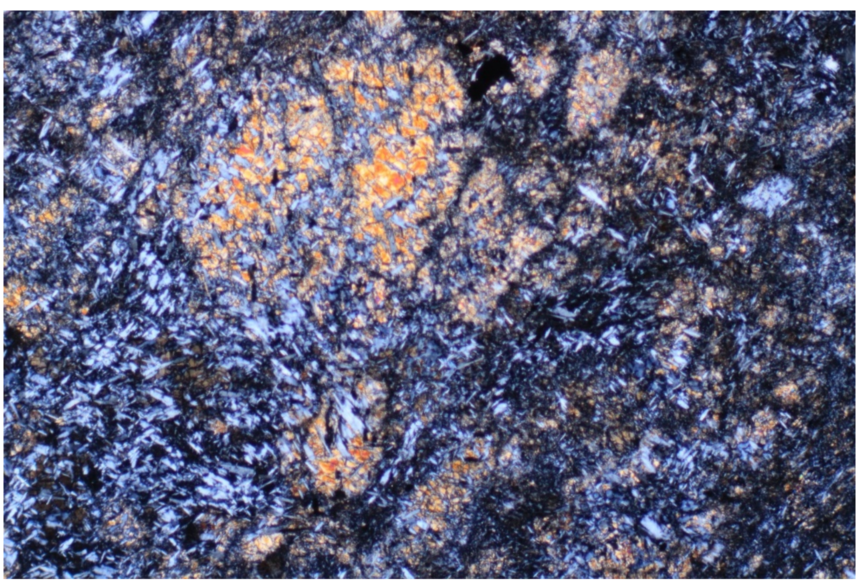


Figure 9-6. Photomicrograph of a serpentinite sample from the stream south of the East Dover Fire Station. The orange mineral is remnant olivine. Note that it is highly fractured with bluish serpentine along the fractures. Field of view is 5 mm.

Other locations in the Appalachians provide much more extensive sampling of oceanic crust that was uplifted and exposed along the Laurentian continental margin. For example, the Bay of Islands complex of Newfoundland provide a near complete section of oceanic crust from uppermost pillow lavas and sediments down to ultramafic rocks. This sequence from pillow basalts to sheeted dikes, to layered gabbros, to dunites, harzburgites and lherzolites, is called an ophiolite sequence. The ultramafic rocks at Dover are very limited in rock types, probably representing a dunitic fragment of what was originally a more complete section. As mentioned above, orogenic events are not delicate.

10. Waits River Formation, Connecticut Valley – Gaspe Trough, South Royalton, VT

In 1759 during the French and Indian War, Robert Rogers lead an attack on the Abenaki village of St. Francis in what is now the province of Québec, Canada. The village was burned and Rogers' Rangers fled southward, some travelling along the Connecticut River that forms the current boundary between Vermont and New Hampshire. Several of the Rangers were captured by the Abenaki, others died of starvation during their efforts to reach the rendezvous location near Fort Number 4 (present day Charlestown, NH). One of the Rangers named Capt. Joseph Wait killed a deer at the junction of Waits and the Connecticut Rivers, leaving a portion for his colleagues. He carved his name on a tree, and eventually, "Wait's" River was named after him. The interesting metamorphosed sediments of this stop share the name Waits, being called the Waits River Formation from its type section along the river. A type section is the originally described location of the formation that serves as the standard to which other portions of the unit are compared.

After the Taconic Orogeny, the eastern margin of Laurentia hosted a large mountain range. With the demise of compression once the island arc was emplaced to Laurentia, the elevated crust began to collapse, creating a basin as the range pulled apart. That basin was the depositional site of sediments that were eroded from Taconic highlands (Figure O-9). The Waits River and the Gile Mountain formations are two sedimentary sheets that were shed into what is referred to as the Connecticut Valley –Gaspe trough.

These sediments were deposited after the Taconic Orogeny, but they are now metasediments, having been metamorphosed during the Acadian Orogeny. After the collision of the island arc and the development of the Connecticut Valley – Gaspe trough, another collisional event occurred along the eastern margin of Laurentia. This time, the culprit was the Avalonia, a microcontinent that collided with Laurentia about 410 to 390 million years ago. This terrane is named after well exposed rocks in Newfoundland.

Both the Waits River and Gile Mountain formations are included in this guide because the initial compositions of the sediments of each formation were different. The Waits River Formation was more carbonate-rich, now represented by marble in the formation. The Gile Mountain Formation was originally more clay-rich, leading to different metamorphic rocks than the Waits River.

Driving Directions

From South Royalton, take RT 110 and cross the White River. Immediately after crossing the bridge, turn south (right) at intersection of RT 110 and RT 14. Follow RT 14 south for 1.5 miles to Ayery's Rocks (Figure 10-1). Park at the paved pull off on right. Or from I-89, take Exit 3 for 0.6 miles on RT 107. Turn south (right) for 5.0 miles south of the RT 107 and RT 14 junction.

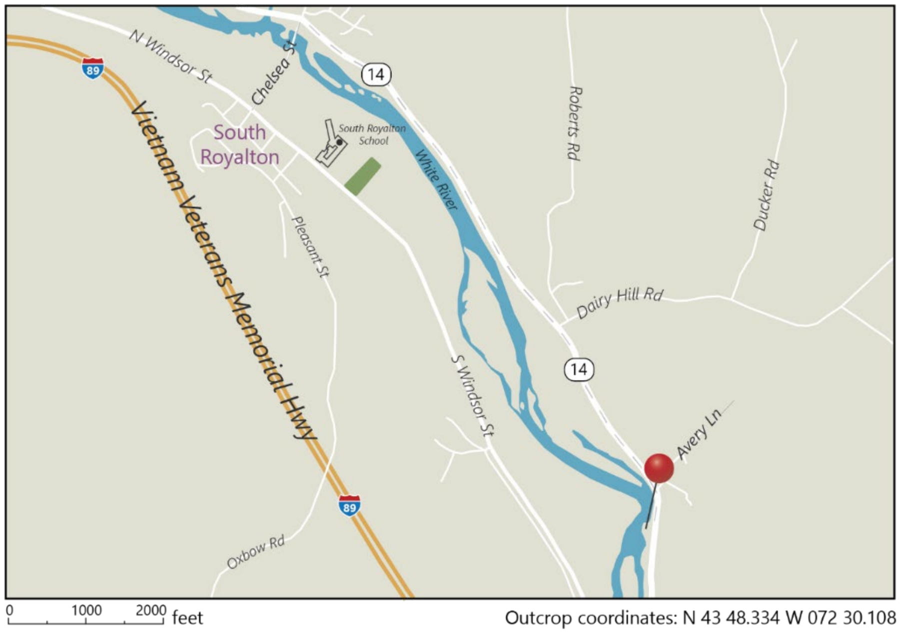


Figure 10-1. Map of the South Royalton area showing the location of Avery's Rocks (red pin).

Figure 10-2. General view of the outcrops of interest along the bank of the White River.

Walking Directions

From pull off, walk west to outcrops along river bank (Figure 10-2; N43° 48.334', W 072° 30.108').

On the Outcrop

These metamorphic rocks are a wonderful mixture of what was originally clay/sand rich portions and carbonate. The clastic sedimentary rocks are now schists containing micas and garnet (Figure 10-3), the limestones are now marbles (Figure 10-4). A very interesting aspect of this outcrop is the difference in flow structures seen in the marble versus the schists. The marble was much more susceptible to flow. Note how contorted the layering is in the marble compared to the relatively intact schist. Also note the broken quartz blocks that are floating in the marble. The quartz blocks were probably quartz veins that were more rigid and were broken during flow of the marble.

Figure 10-3. This photo shows the metasediments that were initially more clay-rich, now having abundant metamorphic garnet.

Figure 10-4. This photo shows the marble in the Waits River Formation. Note the high degree of contortion of the marble.

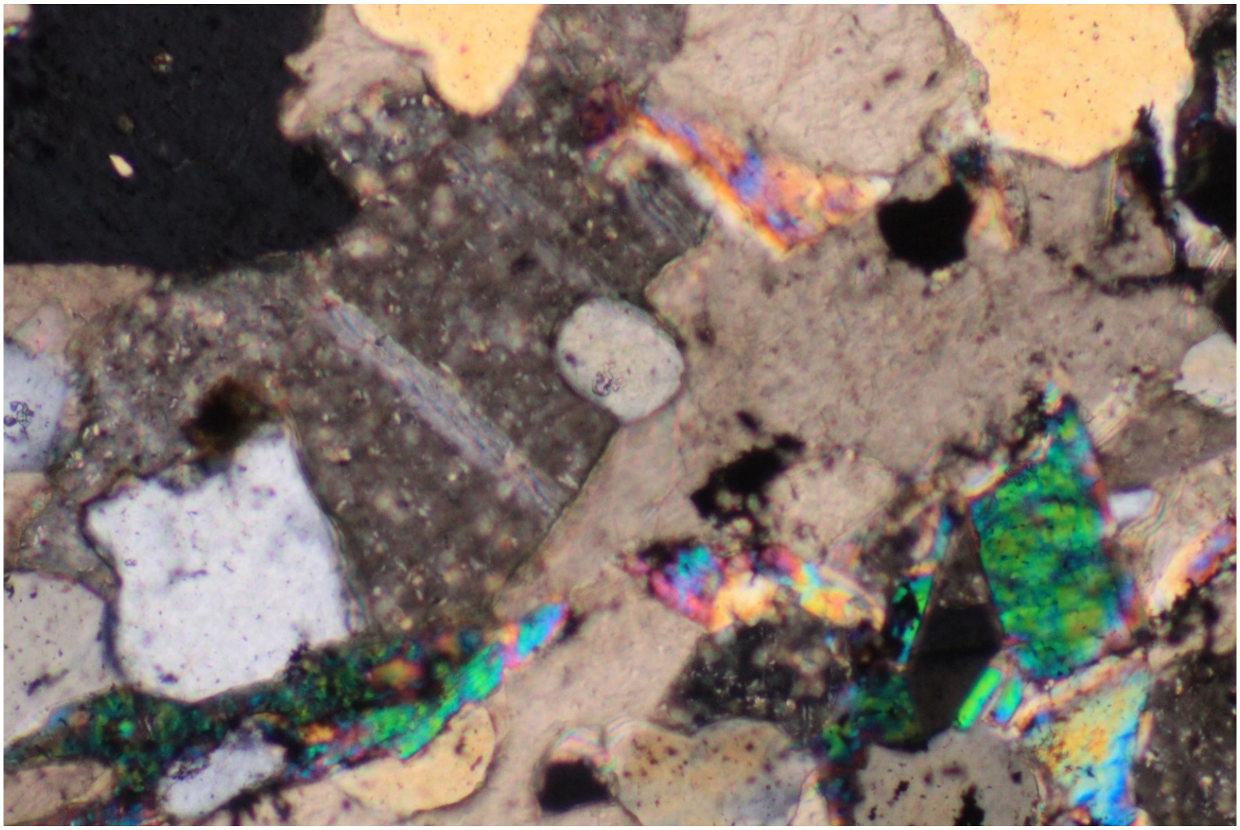


Figure 10-5. Crossed polarized light photomicrograph of the marble of the Waits River Formation. The gray mineral is calcite, the dominant mineral in marble. The green mineral is muscovite. Minor quartz (light yellow) is present at the top of the photo. Field of view is 5 mm.

The presence of garnet and other minerals not shown here (e.g., kyanite) in the clay-rich portions of the formations, and minerals such as actinolite, hornblende, zoisite, diopside, wollastonite in the marble, indicate that the formation has been metamorphosed to amphibolite facies conditions (Figure I-6), reaching temperatures of ~ 550º C and depths of ~ 10-15 km. The loading and heating of the sediments occurred during the Acadian Orogeny resulting from the collision of Avalonia with the Laurentian margin. As noted with the thrust sheets of the Taconic Orogeny of Stops 1, 6 and 7, thrust sheets associated with the Acadian Orogeny were also pushed to the west, loading and metamorphosing the rocks of the Connecticut Valley trough as seen at this stop.

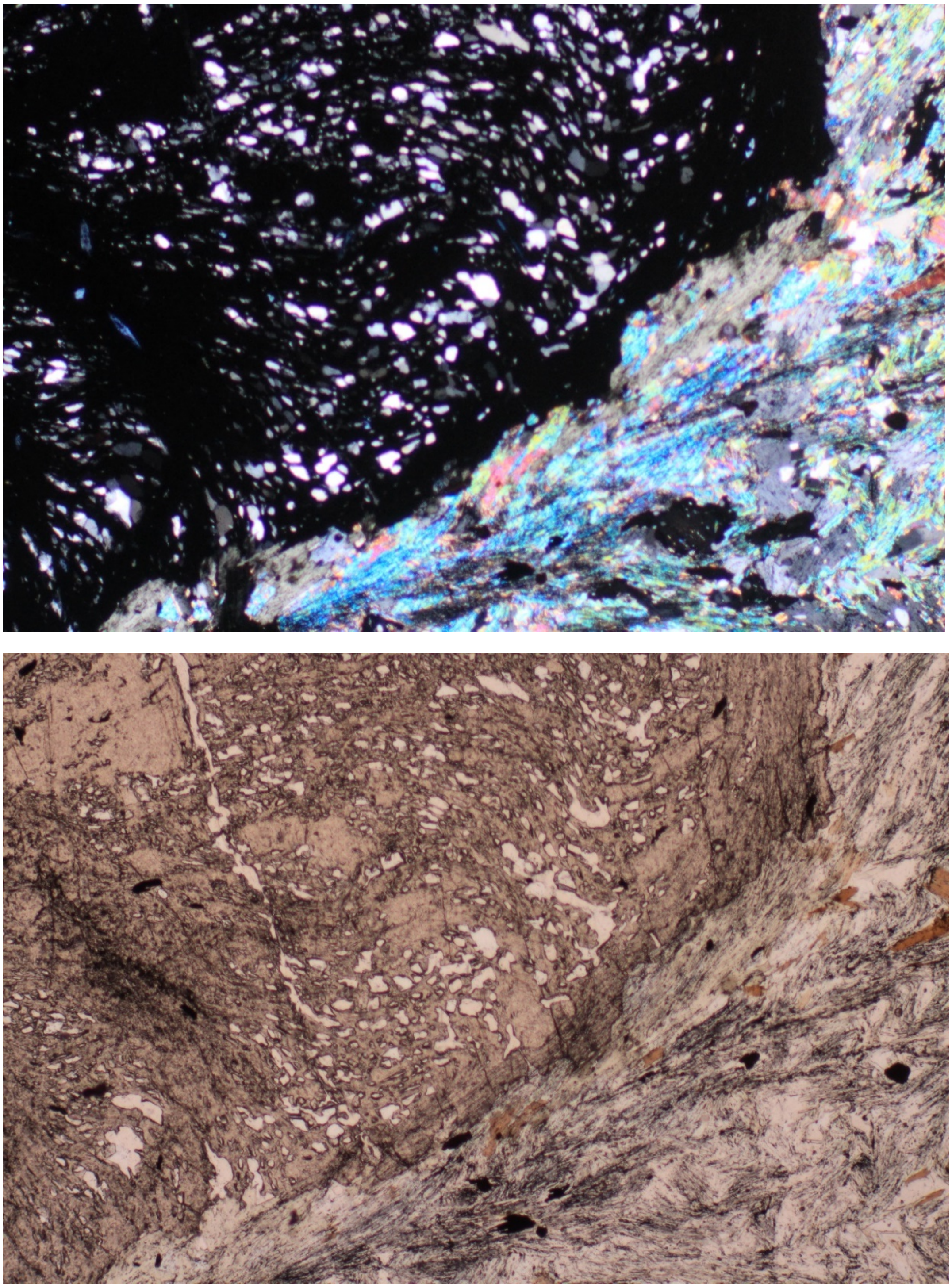


Figure 10-6. Crossed polarized (top photo) and plane polarized light (bottom photo) photomicrographs of the schistose portion of the Waits River Formation. The black mineral of the upper photo is garnet. It contains many inclusions of white, quartz grains that are aligned from an earlier event of metamorphic fabric development and are at an angle to the later fabric defined by the muscovite (bluish-green mineral). Bottom photo is in plane polarized light. Length of photos is 5 mm.

11. Gile Mountain Formation, Connecticut Valley – Gaspe Trough, South Royalton, VT

The Gile Mountain Formation represent clay-rich sediments that were shed into a basin or trough that developed from the collapse of the Taconic Mountains (Figure O-9). This trough, named the Connecticut Valley – Gaspe trough, extended from central New England to Gaspe, Canada. The sediments were subsequently metamorphosed during the Acadian Orogeny about 400 million years ago.

This outcrop along the Connecticut River shows the effects of sediment composition on the metamorphic minerals that formed during the Acadian Orogeny. Some layers were sandy, others more clay-rich. The clay-rich layers host a plethora of interesting metamorphic minerals with garnet being the most obvious. The sandy layers lack this diversity because of their unsuitable chemistry to form garnet and other metamorphic minerals.

Driving Directions

From South Royalton, VT, drive north on Chelsea Street for 0.2 miles to intersection with RT 14. Turn left toward Royalton. At 1.2 miles, pull off on left. Or from I-89, take Exit 2 at Sharon, VT. West on RT 132 for 0.2 miles, at junction with RT 14, turn north. Continue north to intersection of RTs 110 and 14, follow RT 14 north for 1.0 miles, pull off to the left (Figure 11-1).

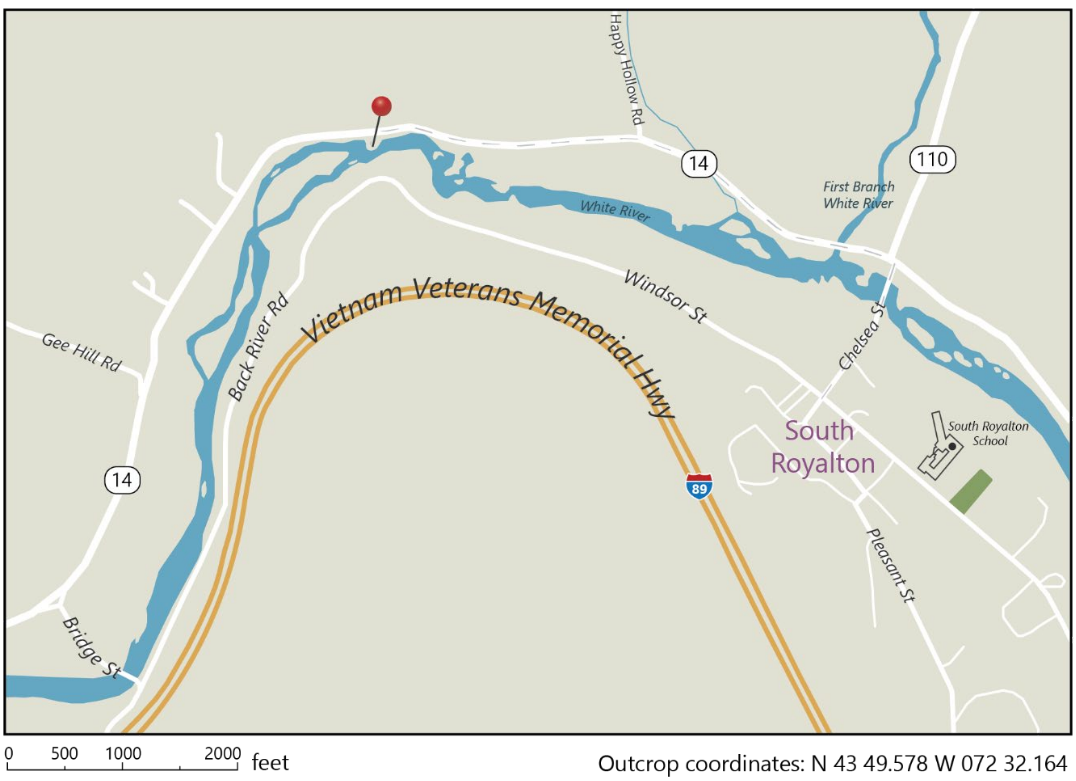


Figure 11-1. Map of the South Royalton, VT area showing the location of the Gile Mountain Formation outcrops along the bank of the White River.

Figure 11-2. General view of the Gile Mountain Formation of this stop.

Walking Directions

From pull off, walk south to a trail that leads down to the river. Outcrops are exposed along the river bank (Figure 11-2; N43º 49.578', W 072º 32.164').

Figure 11-3. Photo showing the variations in the original bedding of the Gile Mountain Formation.

On the Outcrop

The light colored layers to the left of the photo of Figure 11-3 were originally sand-rich compared to the darker, originally clay-rich layers under the hammer. The clay-rich portions are Al-rich, and are much more reactive during metamorphism than sand-rich layers. The dark layers now contain abundant garnet (Figure 11-4), having been metamorphosed during the Acadian Orogeny during the collision of Avalonia with the Laurentian margin. In contrast, the sandly layers are rather monotonous, consisting mainly of quartz and feldspars, lacking interesting metamorphic minerals.

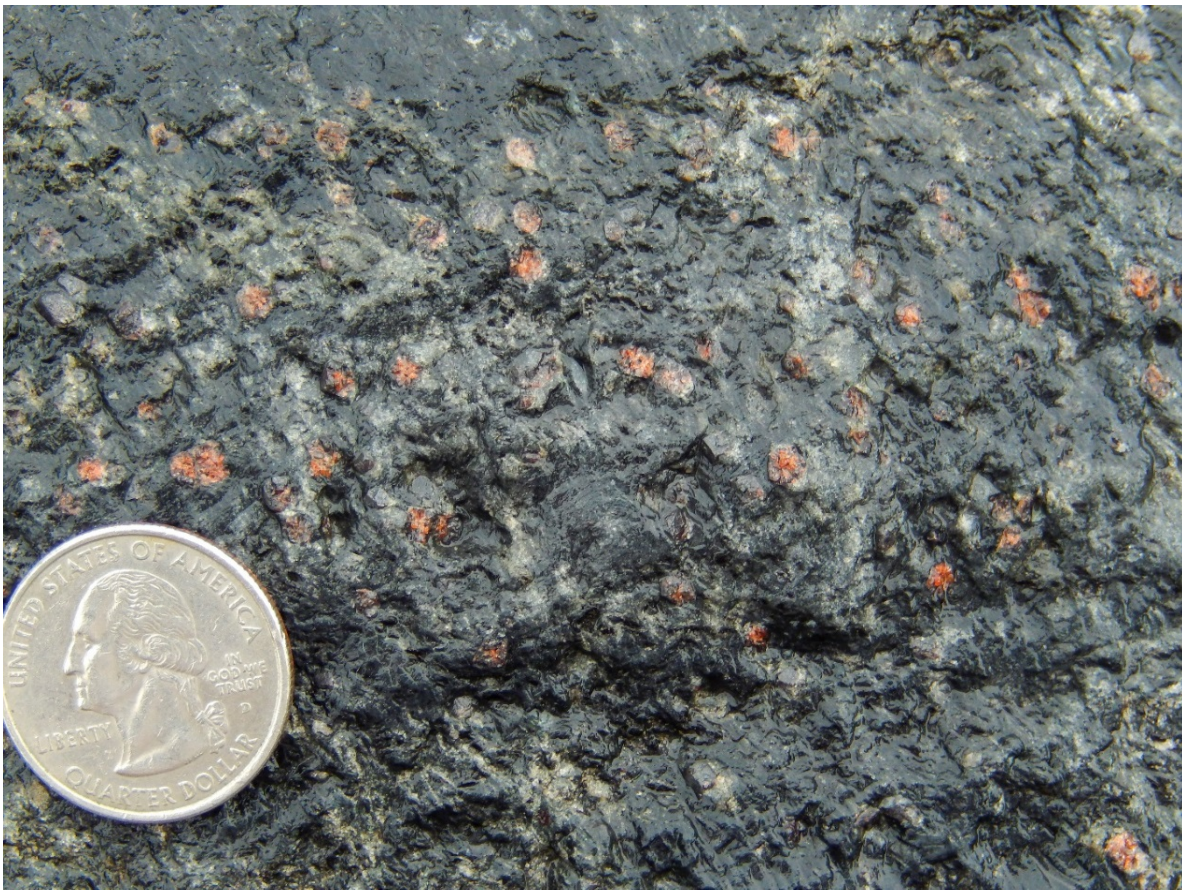


Figure 11-4. This photo, taken of a wet outcrop, shows abundant, reddish colored garnet.

The muscovite of shown in Figure 11-5 experienced two deformational events, the first forming a fabric that was later folded. The lower, plane light photomicrograph shows a wave-like texture to the rock with the initial fabric being aligned approximately from lower left to upper right. That fabric was deformed, forming a younger fabric extending from lower right to upper left. This two fabric texture is called crenulation cleavage.

The Gile Mountain Formation was metamorphosed to amphibolite facies conditions (Figure I-6). During the Acadian Orogeny, the collision of Avalon with the Laurentian margin thickened the crust, causing the sediments to the Gile Mountain Formation to be heated to about ~ 1000º F and buried to depths of ~ 5-10 miles.

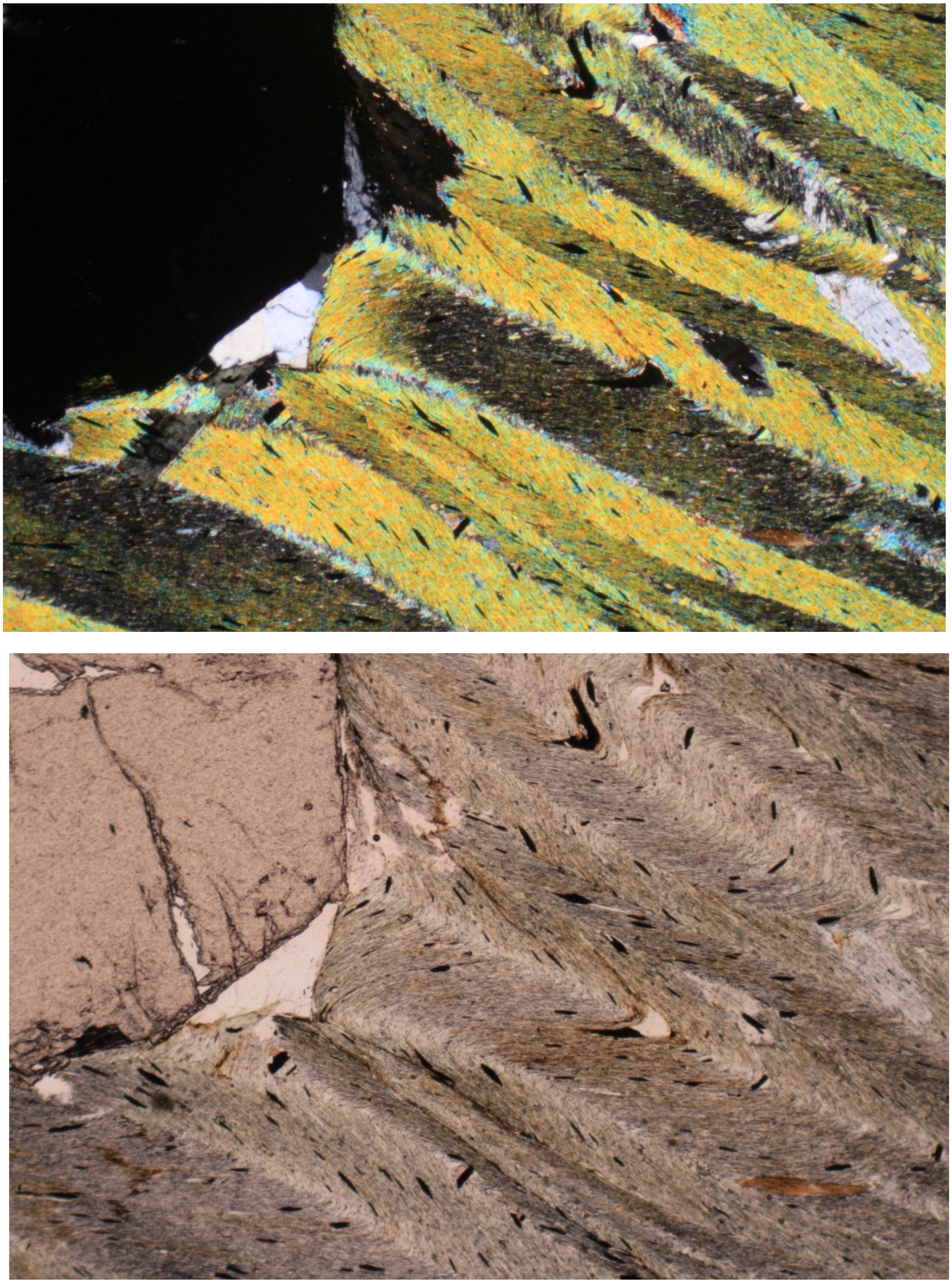


Figure 11-5. Crossed polarized (upper photo) and plane polarized light (lower photo) photomicrographs of the Giles Mountain Formation. The black mineral in the upper photo is garnet. The bands of yellow to dark gray minerals extending from the garnet to the lower right portion of the photo is muscovite.

SUMMARY OF NEW HAMPSHIRE GEOLOGY

The story of New England geology picks up in New Hampshire where Vermont left off. Vermont preserves excellent evidence for the Taconic Orogeny that resulted from the collision of an island arc with the Laurentian margin. Interlaying rocks were deformed, metamorphosed, and trust to the west. Mention was made of this mysterious island arc whose collision pushed the thrust sheets: we can visit volcanic rocks of that island arc in western New Hampshire.

The ~ 470 Ma Ammonoosuc Volcanics represent volcanic rocks of that arc (Stop 12). Some of these volcanic rocks preserve their volcanological features, but most have been subsequently deformed and metamorphosed to amphibolite facies (Figure I-6) conditions during the younger Acadian Orogeny.

Many geologists think that the eastward dipping subduction zone that produced the Taconic island arc then flipped polarity to the west, subducting under the Laurentian margin. A new continental arc formed as the ~ 450 Ma Oliverian Plutonic Suite (Stop 13). These are granitic plutons similar to the Sierra Nevada plutonic rocks of California.

At approximately 430 Ma, a microcontinent called Ganderia collided with Laurentia during the Salinic Orogeny, though the evidence of this orogeny is minor in New England. This was followed by the widespread Acadian Orogeny, the dominant orogenic event that metamorphosed the rocks of northern New England and the cause of granitic magmatism across the region. The Acadian Orogeny, caused by the microcontinent of Avalonia colliding with Laurentia, sandwiched the sediments of the post-Taconic troughs, metamorphosing and melting them. Like the Taconic thrust sheets of Vermont, numerous thrust sheets were pushed to the west during the Acadian collision (Stop 14). The rocks of the thrust sheets reached much higher metamorphic conditions that the Taconic thrust sheets, attaining amphibolite and granulite facies (Figure I-6) because they were originally deeper than most of the rocks of Vermont's thrusts.

Trending from the southwest to the east-central portion of the state is the Central New Hampshire Anticlinorium (Figure NHS-1). This trace across New Hampshire is significant because the rocks of the Central Maine trough were squeezed during the Acadian Orogeny and were pushed upward along this trend. West of the line, the rocks were thrust to the west but east of it, the thrusts moved to the east. This structure is called a pop up or dorsal zone where the thrusts on each side of the zone are essentially mirror images of each other (Figure NHS-2).

Granitic magmas were squeezed upward during the collision and served as lubricants for thrust sheets that overrode them. These magmas formed plutons that were smeared out by the overriding thrusts, some were completely solidified prior to the termination of thrust movement (Stop 16), others were still partially molten (Stop 17). These thrusts and plutons migrated to the west of the pop up zone. In contrast, the plutonic rocks of Stop 18 were emplaced with thrusts that moved eastward of the zone (Figure NHS-2).

The metasediments of central New Hampshire experienced high grade metamorphism at this time, some of which partially melted (Stop 15). A few tens of million years after the Acadian collision, the thickened continental crust melted, producing the Concord-type granites that were emplaced across New Hampshire in into neighboring Vermont and Maine (Stop 19).

Two additional collisional events occurred in the northern Appalachians. The microcontinent Meguma collided about 355 Ma, causing the Neoacadian Orogeny. This was finally followed by the continent to continent collision of Gondwana joining Laurentia to form the supercontinent Pangea in the Alleghanian Orogeny at about 300 Ma (Figure O-13). Rocks of the Ganderian basement of New Hampshire were melted during the Alleghanian and pushed upward, now exposed as the Massabesic Gneiss Complex (Stop 21).

About 200 million years ago, the supercontinent Pangea began to split apart with the first traces of the Atlantic Ocean basin forming (Figure O-14). Tremendous volumes of basaltic magmas were erupted at this time, forming the Central Atlantic Magmatic Province, one of Earth's largest flood basalt provinces (Figure O-15). No lava flows remain in northern New England, but the dikes that feed the lava flows are present in eastern New Hampshire (Stop 22) and coastal Maine (Stop 42).

Finally, the syenites, quartz syenites and anorogenic granites of the White Mountain Volcanic and Plutonic Suite were emplaced between 200 and 100 million years ago (Stops 23-26). These form a linear belt of plutons that extend in a northerly direction across New Hampshire, linking the New England Seamounts with the intrusions of the Monteregian Hills of Québec (Figure O-16).

Simplified Geologic Map of New Hampshire

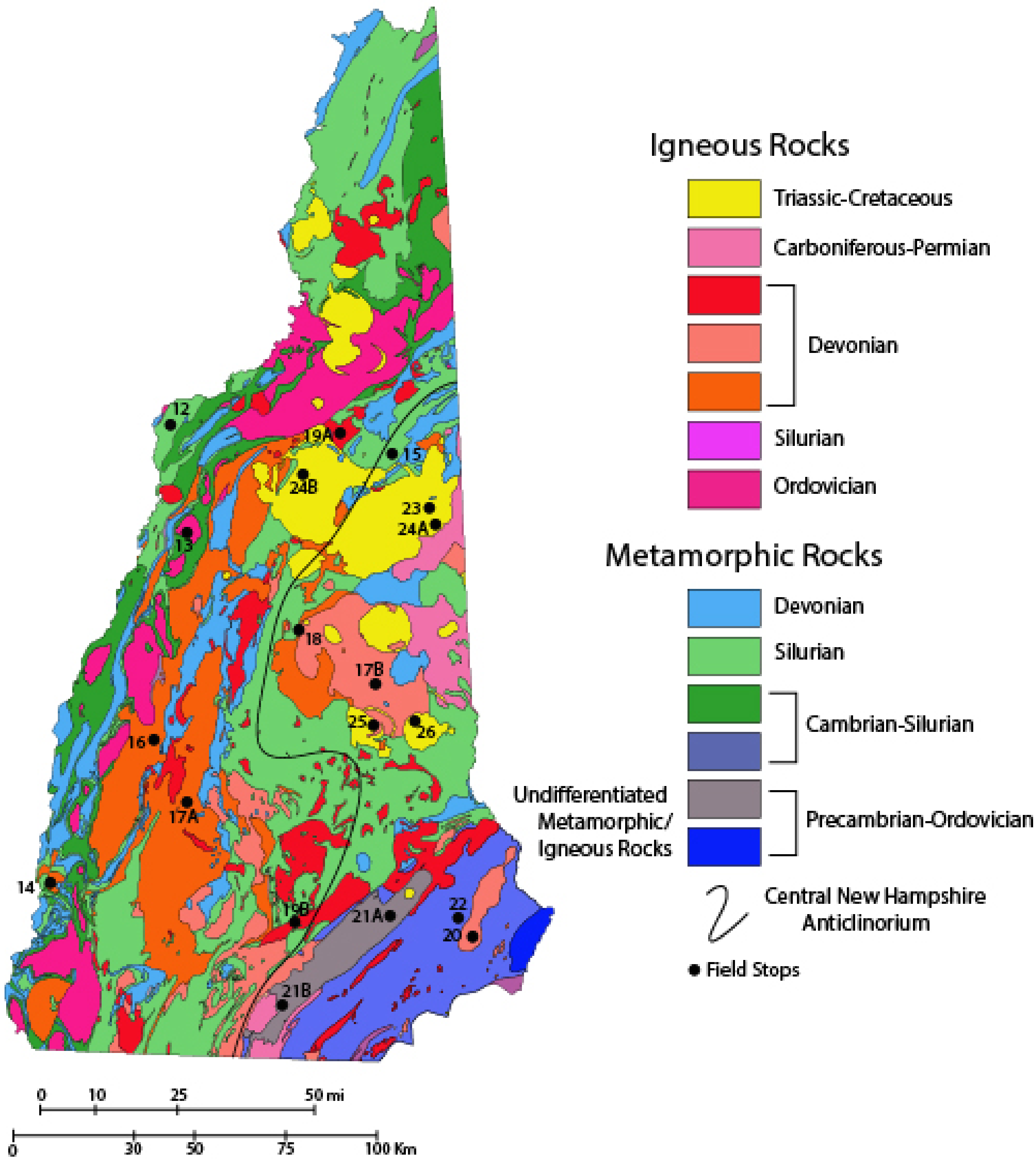


Figure NHS-1. Simplified geologic map of New Hampshire (after Lyons et al., 1997).

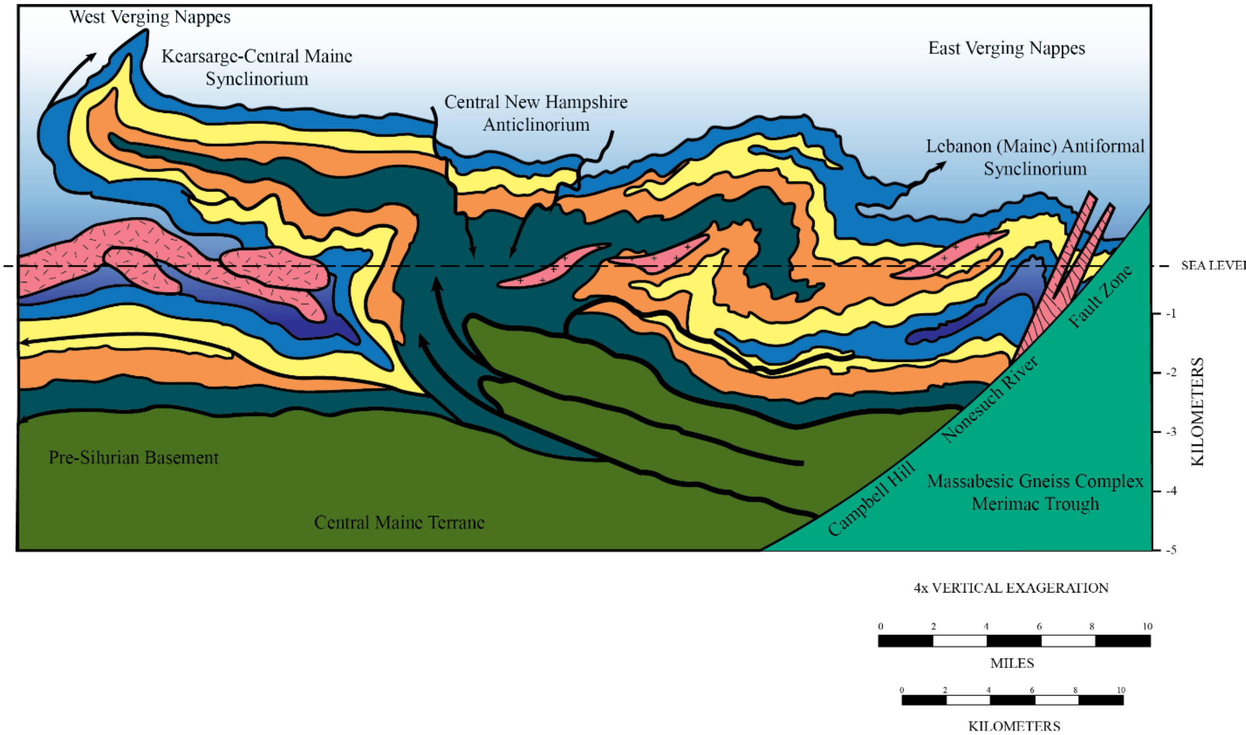


Figure NHS-2. Cross section of NH showing the Central New Hampshire Anticlinorium and the trust faults and nappes that extend both to the west and to the east. Plutonic rocks form the base of the nappe to the west (After Eusden and Lyons, 1993).

12. Ammonoosuc Volcanics, Ordovician Island Arc, Lisbon, NH

In several descriptions of the field stops in Vermont, the collision of a mysterious island arc with Laurentia was mentioned as the cause of the Taconic Orogeny. Where is that island arc now? It is well exposed at this and other locations in northern New Hampshire and eastern Vermont.

Picture in your mind an ancient volcanic arc similar to the Japanese or the Philippines Islands. These modern arcs developed as the Pacific Oceanic plate subducted beneath another oceanic plate, producing a series of volcanoes that grew to form a chain of volcanic islands. The Ammonoosuc Volcanics erupted from an island arc that formed in a similar manner: an oceanic plate subducted eastwardly under another oceanic plate to form an island arc between 480 and 450 million years ago (Figure O-5). Eventually, the eastward subducting oceanic plate was completely subducted and the ocean basin between the island arc and the Laurentian continent was consumed. This led to the edge of the continent partially entering the subduction zone, lifting the island arc and thrusting it over the edge of the Laurentian margin (Figure O-7). This collision created the thrust sheets (Stops 6 and 7), the emplacement of ultramafic rocks (Stop 10), and mountains associated with the Taconic Orogeny.

Driving Directions

From center of Lisbon, NH, drive north on RT 10 (also marked as RT 302; Figure 12-1). At 0.6 miles north of the RT 10 and RT 117 junction, there are good road cuts on the left (west; Figure 12-2). Pull off along the side of road. If you are driving north on I-93, take Exit 42 at Littleton, NH. Starting where the off ramp joins RT 10, turn left and drive south for 6.6 miles. The road cuts are along the right side of the road.

Figure 12-1. Map showing roadcut of Ammonoosuc Volcanics along RT 10, northeast of Lisbon.

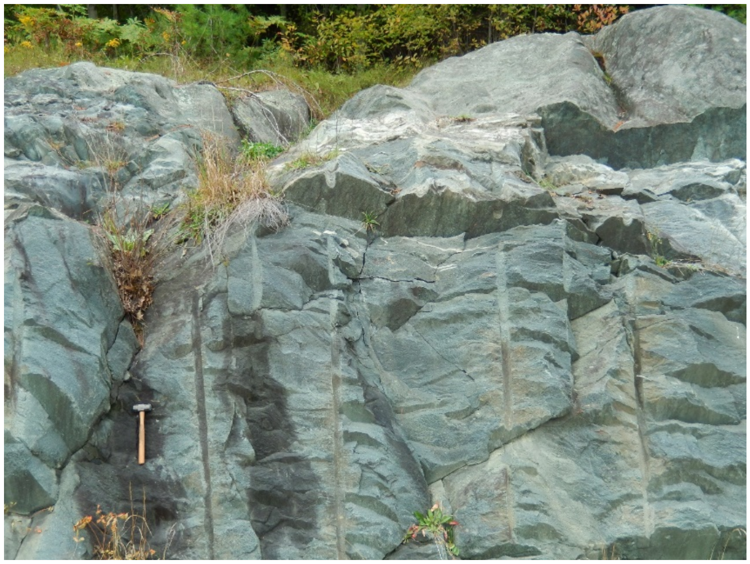


Figure 12-2. General view of outcrop of Stop 12 (N44º14.618', W071º52.430').

On the Outcrop

Most of the rock at this road cut is metamorphosed basalt. No original volcanic features such as pillow basalts are evident at this location (though there are original features preserved elsewhere). The rock composition is still very similar to the original basalt, but it has been subjected to higher temperature and pressure conditions during the Taconic Orogeny and probably the younger Acadian Orogeny to form amphibole and plagioclase feldspar, indicative of amphibolite facies metamorphism (Figure I-6). Analyses of these rocks indicates basalt compositions that are similar to modern-day island arcs.

What makes this outcrop especially interesting is the elongated needles of amphibole (Figure 12-3). The amphibole crystals are concentrated along foliation planes but within those planes, they show no consistent orientation.

Figure 12-3. Photo of Ammonoosuc Volcanics showing needle-like amphibole crystals

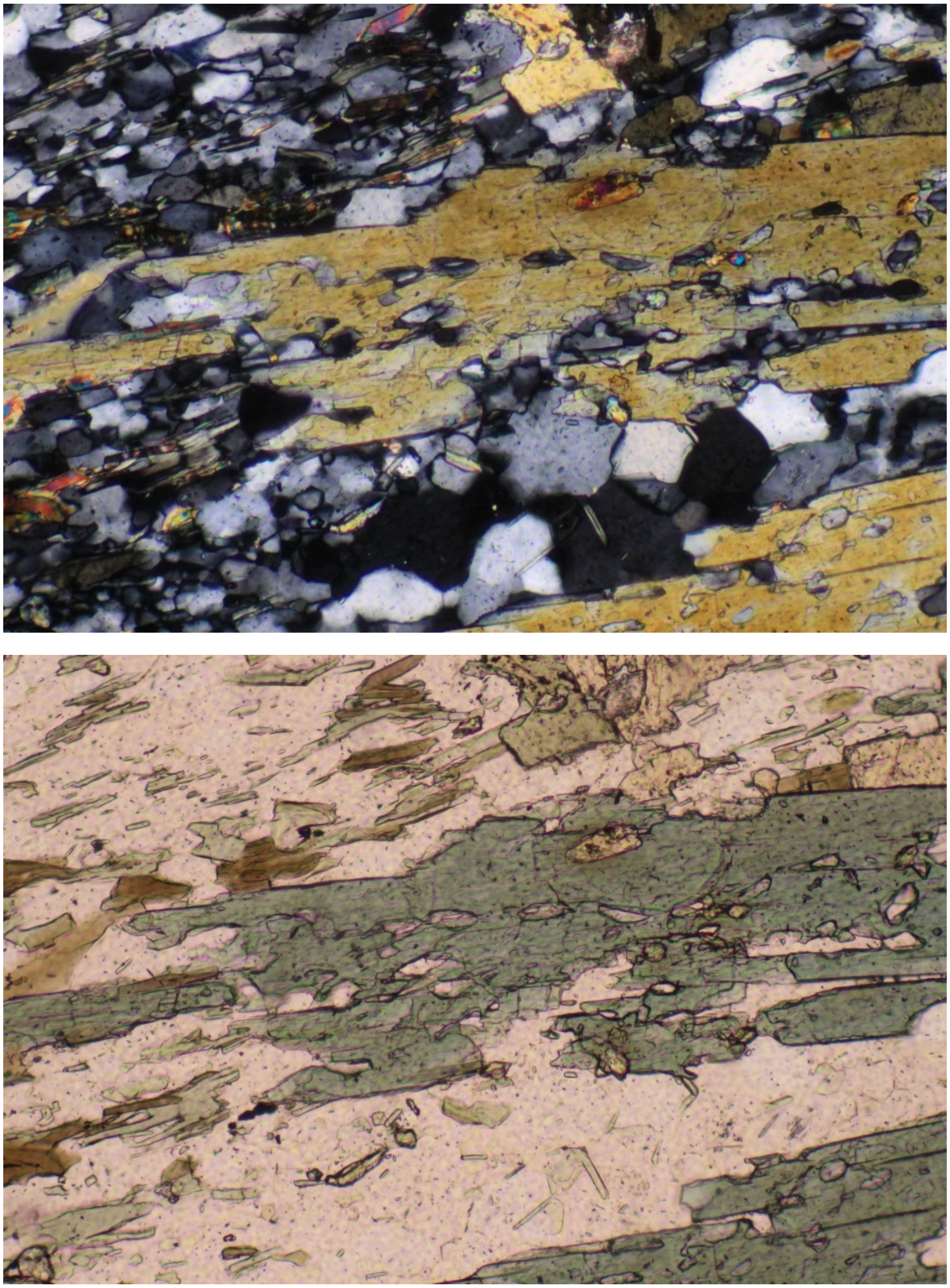


Figure 12-4. Crossed polarized (upper photo) and plane polarized (lower photo) photomicrographs of the Ammonoosuc Volcanics. The yellowish mineral of the upper photo is hornblende, a type of amphibole (green mineral of the lower photo). Much of the remainder of the rock is plagioclase feldspar. Minor biotite (brown mineral of lower photo) is present. Note the alignment of the amphibole grains that occurred during metamorphic deformation.

Figure 12-5. Felsic metavolcanic rocks of the Ammonoosuc Formation.

Felsic metavolcanic rocks are present at the northern end of the road cut. These felsic rocks erupted in between basaltic flows. As with the surrounding amphibolites, the felsic volcanic rocks have been metamorphosed and deformed.

13. Owls Head Pluton, Oliverian Plutonic Suite, between Haverhill and North Haverhill, NH

The Oliverian plutons are granitic rocks that intruded into the Taconic-modified margin of Laurentia. Like the Ammonoosuc Volcancis, they are also related to arc magmatism associated with subduction, but tell a different story than the Ammonoosuc Volcanics. The Ammonoosuc Volcanics were erupted in an island arc, similar to the basaltic rocks forming in the modern arcs in Japan or the Philippines. In contrast, the Oliverian plutons are more granitic in composition, suggestive of an arc built on continental crust.

Two conflicting models have been proposed to explain the change from island arc to continental arc magmatism. One is that the eastward subduction of the oceanic plate ceased when the continental margin started to subduct. Continental crust is less dense than oceanic crust and because of its buoyancy, it didn't subduct very deep before shutting down the system. The connection between the subducted oceanic crust and the trailing continental margin may have severed, allowing the oceanic crust to sink into the mantle (Figure O-7). Subsequent compression could have initiated a new subduction zone, this time dipping to the west under the Laurentian continental margin. As the subduction-produced magmas ascended through continental crust, they evolved to more feldspar- and quartz-rich compositions, indicative of a continental arc setting. An alternative model is that subduction was continuous to the east and the arc magmas ascended through a small fragment of Laurentian crust called Dashwoods. In either case, the Oliverian plutons have compositions that indicate a continental arc, showing the transformation of island arc off the continental margin to a continental arc similar to the Sierra Nevada batholith of the western United States.

One major difference between the Sierra Nevada batholith and the Oliverian plutons is that the Oliverian Plutons were metamorphosed and deformed in subsequent orogenic events, having been domed upward during younger orogenies. Acadian and Neoacadian metamorphism has modified the rocks so they no longer preserving their original igneous textures and mineral compositions. The rocks of the Owls Head Pluton of this stop is one of many that form a series of Oliverian domes that extend from Long Island Sound in Connecticut to northern New England; the domes of New Hampshire are shown in Figure 13-1.

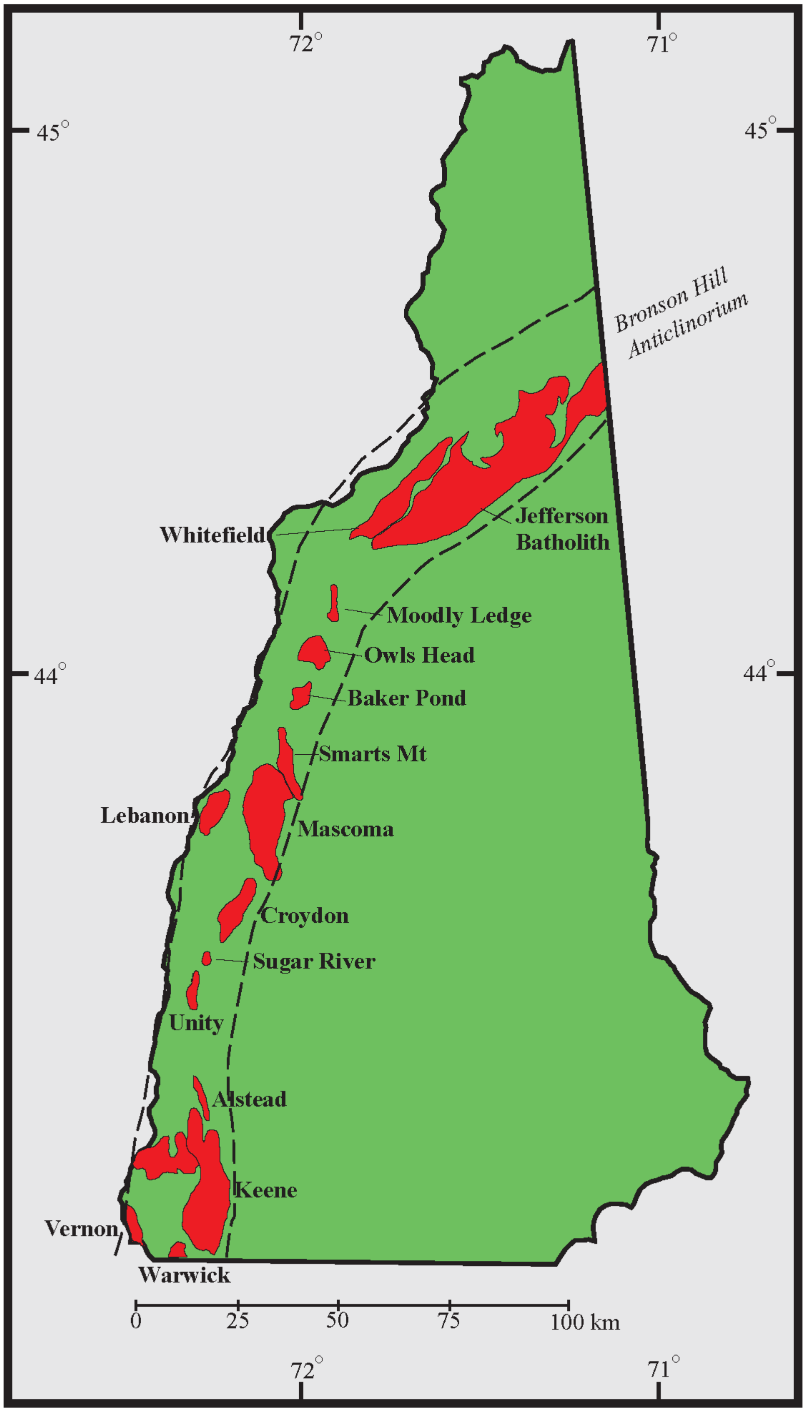


Figure 13-1. Generalized map showing the locations of Oliverian Domes in New Hampshire (After Dorais et al., 2008).

Driving Directions

Drive north of Haverhill, NH, on RT 10 to junction of RT 10 and RT 25. Turn east (right) on RT 25 and drive for 7.5 miles (Figure 13-2). Pull off on left at the Baker River Flood Control sign at Oliverian Pond.

Figure 13-2. Map of the Oliverian Pond area showing the cuts of interest for this stop.

Figure 13-3. General view across Oliverian Pond to the cliffs of Owls Head.

Walking Directions

Walk a few hundred yards to the northeast across the dam to the large cuts on the north side of Oliverian Pond (Figure 13-4).

Figure 13-4. General view of cuts at north side of the dam (N44º00.371', W071º55.703').

On the Outcrop

Figure 13-5 shows a weak alignment of the minerals in the Owls Head Pluton. The orientation of biotite is the easiest to see; it is aligned in subhorizontal streaks (Figure 13-6). The feldspars, both Kspar and plagioclase, and quartz are equant, rather stubby shaped minerals and are not as susceptible to realignment. Nonetheless, a weak subhorizonal orientation is evident for these minerals as well. The degree of development of this fabric in the Oliverian rocks is strongest near the contacts between plutons and the overlying Ammonoosuc Volcanics, and very weak to lacking in the cores of the domes. Many geologists think this fabric was developed during the Neoacadian Orogeny when the granitic plutons of the Oliverian Suite were uplifted, causing the overlying Ammonoosuc Volcanics to slide downward off the domes.

Figure 13-5. Photo of the Oliverian rocks of the Owls Head Pluton. Note the gneissic fabric of the rock.

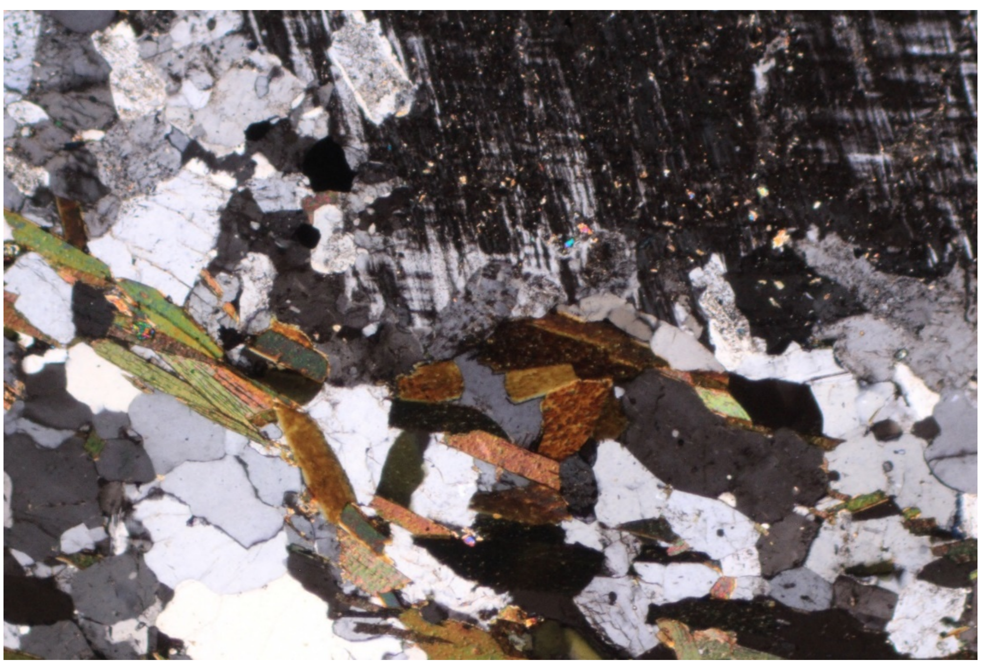


Figure 13-6. This crossed polarized light photomicrograph shows a microscopic view of the minerals of the Owls Head Pluton. The green and brown elongated minerals in the lower portion of the photo are biotite. Note the alignment of the micas, forming a foliation in the rock. The large crystal in the upper right is microcline, a form of Kspar. Many of the smaller gray to white crystals are quartz which has undergone grain size reduction during deformation.

Figure 13-7. Mafic dike cutting the Oliverian, Owls Head Pluton (N44°00.359', W071°55.676').

Another interesting feature at this location is the presence of a mafic dike that cuts up through the gneiss (Figure 13-7). No chemical analyses are available for this dike, making it difficult to place it into a geologic setting, but it cuts the ~450 million year old Oliverian gneiss, constraining the dike to be younger than that age. The fact that the dike is metamorphosed indicates that this mafic magma was not related to the breakup of Pangea as are the younger, unmetamorphosed dikes described at Stops 22, 35c and 42. It is probably post Taconic but pre-Acadian.

14. Fall Mountain Thrust of the Acadian Orogeny, Central Maine Trough, North Walpole NH

At Lone Rock Point along Lake Champlain in Vermont, we saw the most famous thrust fault in New England. The carbonates of the Dunham Formation were pushed up and over the shales of the younger Iberville Formation. Very little change in the minerals of either the carbonates or the shales occurred during the thrusting because the rocks were never very deep or hot. Such is not the case here with the Fall Mountain thrust. This thrust sheet was pushed up from far deeper in the crust and magma was injected upward with the thrust sheet, forming a sheet of lubricating, Bethlehem Granodiorite over which the thrust moved (Figure NHS-2). The thrust continued to move even after the Bethlehem completely solidified, severely deforming and metamorphosing the Bethlehem Granodiorite as seen at Stop 16.

The effects of the duel emplacement of the thrust and magma wasn't limited to deformation of the pluton; the pluton also affected the metamorphic rocks of the thrust sheet. Because of the proximity of the Bethlehem Pluton, the Rangeley Formation was metamorphised from an andalusite-bearing rock to higher temperatures of the sillimanite grade at essentially constant pressure. Geologists who have studied these rocks have determined that the rocks experienced an increase of about 300 degrees Fahrenheit because of the contact metamorphic effects of the pluton.

Following the emplacement of the Fall Mountain thrust and the Bethlehem Granodiorite, another thrust sheet called the Chesham Pond thrust was stacked on top of the Fall Mountain thrust sheet. The Kinsman Granodiorite forms the sole of this thrust. As we saw in Vermont, the successively younger thrusts come from deeper in the crust than the sheet that they overrode.

The thrust sheets of New Hampshire originated from the Central New Hampshire Anticlinorium (Figure NHS-1), a zone along the length of New Hampshire. Imagine two snow plows pushing snow into each other's plow. The snow compressed between the plows would be pushed upward. Such were the rocks along the Central New Hampshire Anticlinorium where the thrusts originated at depth but were pushed upward between the "snow plows" of Laurentian and Avalonia during Acadian Orogeny. Most of the thrusts were pushed up and out to the west, including the trust seen here. However, some thrust sheets and their associated magmatic soles moved from the anticlinorium to the east (Figure NHS-2). The Spaulding Tonalite of Stop 18 is one of these.

Driving Directions

From Arch Bridge in North Walpole, NH, drive south on RT 12 for 0.5 miles. Just north of the Green Mountain Railroad area, pull off to the east side of the road and park (Figure 14-1). This is just north of the Vilas Bridge Road that junctions with RT 12. Don't park at the railroad property parking area. Just after 911, I made the mistake of doing so when the feds were very concerned about people trespassing on federal land. I practically had to sell my soul to the devil to get my car released.

Figure 14-1. Map of the North Walpole, NH area showing the location of the outcrops along the bank of the Connecticut River.

Figure 14-2. General view of the outcrops of interest at the Vilas Bridge (N43º08.150', W072º26.400').

On the Outcrop

Rocks along the Connecticut River below the bridge (Figure 14-2) have been metamorphosed to upper amphibolite facies (Figure I-6). These rocks were pushed up from the east and in the process, overrode the granitic magmas below. The Bethlehem Granodiorite (better seen at Stop 16) is located closest to the river with the Rangeley Formation overlying it. The contact is the Fall Mountain Thrust Fault.

Figure 14-3. Photo of a pegatitic pod in the Rangeley Formation.

Figure 14-3 shows a small pegmatic pod consisting of feldspars and quartz. The dark mineral is tourmaline, a mineral that contains about 8% boron. In contrast to sediments deposited in continental basins that are boron poor, boron is enriched in some marine sediments. The sediments that were lithified to form the Rangeley Formation were deposited in the Central Maine trough, a marine basin adjacent to the Laurentian margin.

The metasedimentary rocks at this location are well known because of these large, gray colored crystals. These were originally andalusite (Al_2SiO_5) that forms as a metamorphic mineral at low pressures (Figure I-6). The nearby Bethlehem Granodiorite of the Bellows Falls Pluton (located adjacent to the river) heated the overlying metasediments after the initial metamorphism to the point where andalusite was no longer stable, being replaced by its higher temperature cousin sillimanite (also Al_2SiO_5). Some of these crystals are mantled by muscovite that formed when the rocks were cooling from their temperature peak during the Acadian Orogeny. Some of these altered crystals are susceptible to weathering and erode to form depressions in the rock. The depressions are informally called turkey tracks. You'll easily see many turkey tracks as you look at the tops of the beds at this stop (Figure 14-5).

Figure 14-4. This photo shows abundant garnet in the Rangeley Formation.

Figure 14-5. “Turkey tracks” in the Rangeley Formation.

15. Rangeley Formation Migmatites, Central Maine Trough, Pinkham Notch, NH

Did you ever wonder where granitic magmas form? This outcrop of the Rangeley Formation at this location shows us one origin. The rock here is a migmatite, that is, a high grade metamorphic rock that reached granulite facies (Figure I-6) and has undergone partial melting. Migmatites are mixed rocks with the lighter colored bands representing the granitic melts and the darker bands representing the residue remaining after the melt has been extracted and concentrated. Migmatites such as these seen here are the birthplaces of granitic plutons. This is where the melting occurs, and under suitable circumstances, the melt accumulates and is extracted from the source in dikes that migrate upward in the crust. Multiple dikes coalesce and eventually form larger bodies of granitic plutons.

Driving Directions

At the RT 16 and RT 2 junction in Gorham, NH, drive south on RT 16 for 6.3 miles to the Great Gulf Wilderness Area Parking lot on west side of the road (Figure 15-1). It is contrary to the Live Free or Die mentality of the people of NH to have to pay for parking, but the National Forest Service uses parking fees to maintain the National Forest. The charge at the time of writing is \$5 per day. Follow the parking lot road to the northern end.

Figure 15-1. Map of the Great Gulf Wilderness Area Parking lot south of Gorham, NH.

Walking Directions

From the northern end of the lot, walk for 0.1 miles to the suspension bridge. Good outcrops are under the bridge along the east side of the river.

On the Outcrop

This outcrop shows abundant light colored bands that represent granitic melts (Figure 15-2). These light bands are called leucosomes. The dark, mica-rich portion of the rock is thought to represent partially melted portions of the migmatite, i.e, melt has been extracted from the dark bands, leaving a residue behind. These dark bands are called melanosomes. Cutting across both bands are small granitic dikes. When the melt fraction reaches amounts of ~ 40% of the rock, the melts mobilize to form dikes. These ascend upward in the crust and may feed larger plutons higher in the crust.

Figure 15-2. Migmatite of the Rangeley Formation.

Figure 15-3 shows a block of more homogeneous rock sitting within the migmatite. This block has a different composition than the migmatite, having a lower abundance of mica and more feldspar and quartz. It probably represents a sandy layer in an originally clay-rich rock that was just like the more quartz-rich layers seen in the Gile Mountain Formation of Stop 11. The clay-poor, quartz-rich composition of these layers have a higher melting temperature than the adjacent rock, hence the sandy compositions lack abundant leucosomes found elsewhere in this outcrop. Unmelted portions of a migmatite are called the paleosome. This paleosome was more rigid than the partially melted migmatite and upon shearing, the more rigid layer broke into blocks as seen here. In contrast, the clay-rich compositions melt at lower temperatures than the sandy portions, hence, the abundance of granitic leucosome in the initially, clay-rich portions of the rock.

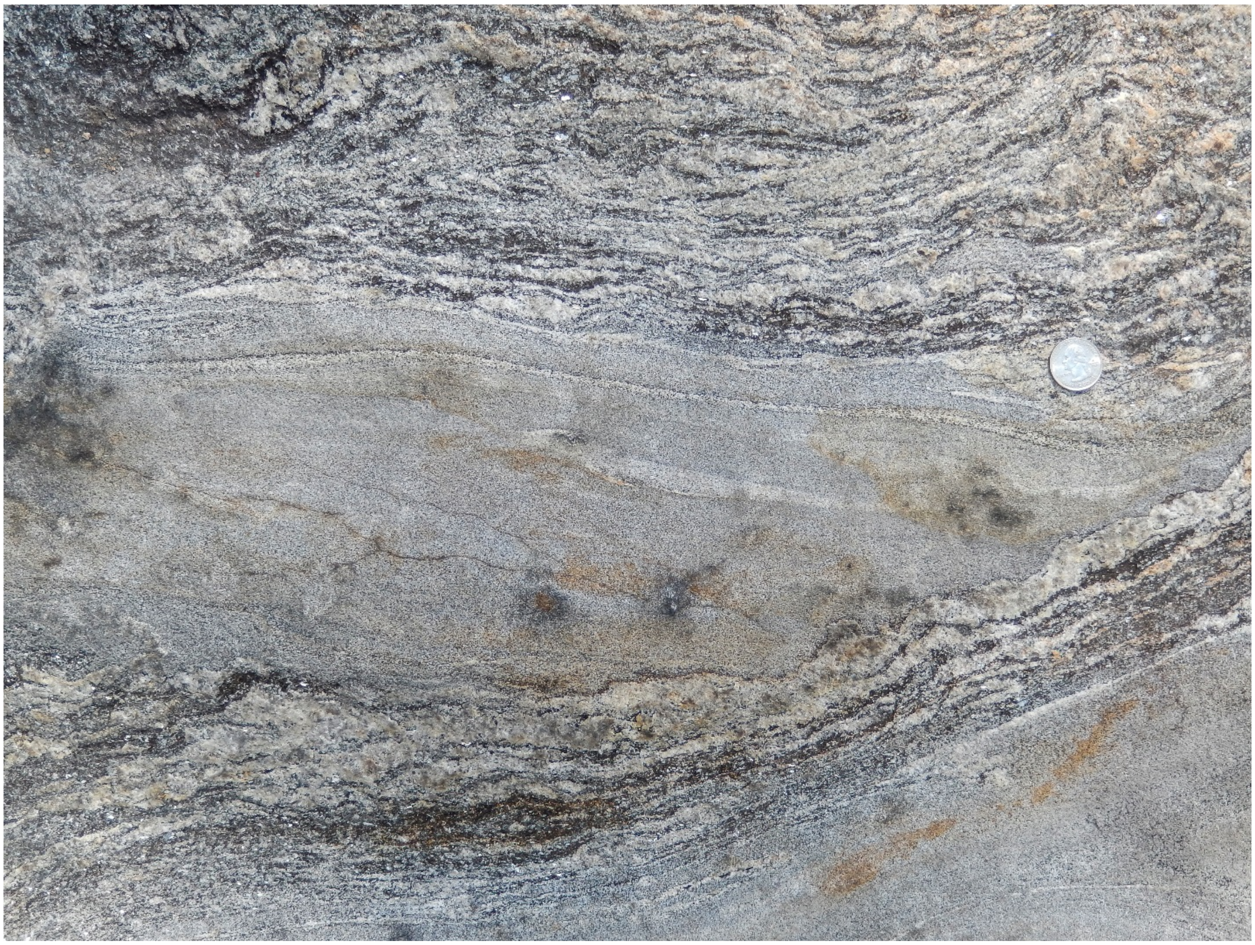


Figure 15-3. Unmelted portions of the Rangeley Formation.

16. Bethlehem Granodiorite, New Hampshire Plutonic Suite, Springfield Rest Stop, I-89, NH

Three members of the New Hampshire Plutonic Suite were emplaced as sheets of magma in between thrust sheets during the Acadian Orogeny. Two of these members, the Bethlehem Granodiorite and the Kinsman Granodiorite, form several elongated plutons extending from southwestern New Hampshire near Keene to the west of the White Mountains in the Franconia Notch region (Figure NHS-1).

The Bethlehem Granodiorite plutons are relatively thin sheets, less than 1 mile thick, that were emplaced during Acadian thrusting that pushed metasedimentary rocks of the Central Maine Trough to the west. The overlying Brennan Hill thrust not only smeared the magma into a thin sheet as it were moving to the west, but the continual motion of the thrust to the west after solidification of the pluton deformed the Bethlehem plutonic rocks in most locations to a metamorphic rock called a gneiss. Traditionally, the Bethlehem Granodiorite was called the Bethlehem Gneiss because of its strong fabric developed in the solid state after crystallization.

Because the pluton was smeared and deformed during Acadian thrusting, determining the age of the pluton reveals when the thrust sheets were moving westward during the Acadian Orogeny. The Bethlehem Granodiorite is ~ 410 million years old, showing that thrusting was of that age and slightly younger in this portion of New Hampshire.

Driving Directions

From Concord, NH, follow I-89 northwest to highway mile marker 39.4 and turn into the Springfield Rest Stop (Figure 16-1). Park at the eastern edge of the lot and walk to the right side of the large cuts (Figure 16-2).

Figure 16-1. Map of the Springfield Rest Stop area along I-89 northwest of Lake Sunapee, NH.

Figure 16-2. General view of cuts of the Bethlehem Granodiorite at the rest stop (N43º28.017’, W072º06.056’).

On the Outcrop

This outcrop of the Bethlehem Granodiorite is located in the Mt. Clough Pluton (Location 16 of Figure NHS-1). Granodiorite is a granitic rock that contains more plagioclase feldspar than K feldspar (Figure I-5).

The Bethlehem Granodiorite and the Kinsman Granodiorite of Stop 16 are very similar in composition. However, they appear very different in the field, partly related to the crystallization state of the two magmas while thrusting was taking place in western New Hampshire. Both the Bethlehem and Kinsman Granodiorites were partially molten when the plutons were emplaced during thrusting. They differ however, in that the thrusting of the Brennan Hill thrust over the Bethlehem continued even after the Bethlehem magmas solidified, causing shearing and deformation of the then solidified pluton.

The rock has a well-developed gneissic fabric (Figure 16-3). Note that the white feldspars tend to have tapered ends where they come to a point. Dark micas (biotite) form elongated clusters, defining a planar pattern. The tapered feldspars and elongated mica clusters form a gneissic fabric, having formed in the solid state and were not produced while the pluton was molten.

Further evidence of post-emplacement deformation of the pluton is shown by this well-developed shear zone (Figure 16-4). A close-up view reveals that the rock has been ground up as the rock in the upper part of the photo moved to the west (to the left) with respect to the rock below the shear.

Figure 16-3. Post crystallization deformation is very evident at this outcrop.

Figure 16-4. Shear zone in the Bethlehem Granodiorite.

The crossed polarized light photomicrograph of Figure 16-5 shows a microscopic view of the mica foliation in the Bethlehem Granodiorite. Mica (blue, yellow, brown and greenish elongate minerals) is susceptible to realignment during shearing, with the mica crystals rotating and re-growing in an aligned orientation from shearing.

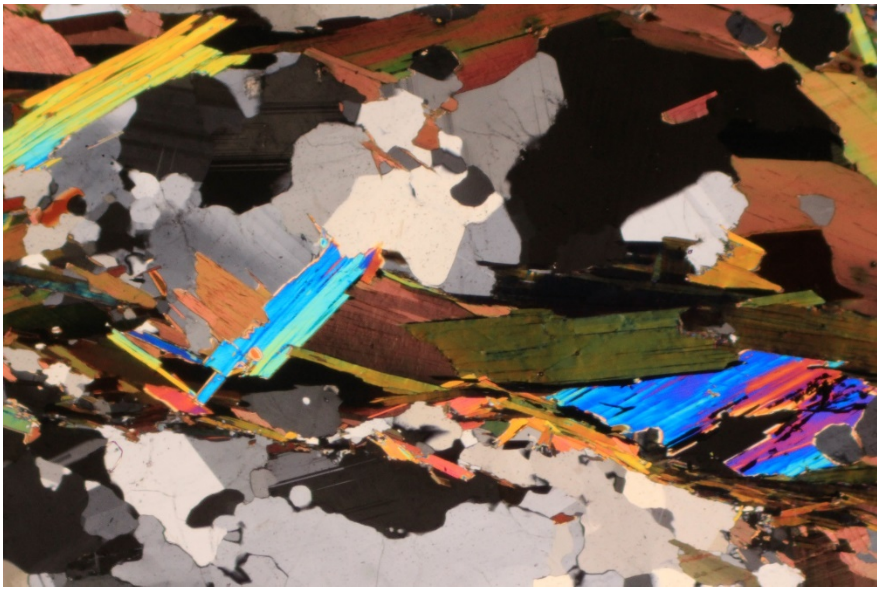


Figure 16-5. Photomicrograph of the Bethlehem Granodiorite. Field of view is 5 mm.

Figure 16-6. Mafic enclave in the Bethlehem Granodiorite.

One more feature to note at this location is the clusters of minerals forming darker rock immersed in the Bethlehem (Figure 16-6). These darker rocks are called xenoliths if they are fragments of the rocks surrounding the pluton that broke off and were floating like rafts in the magma. Other fragments of dark rock are the result of a magma of more mafic composition interacting with the Bethlehem magma. A mafic magma injecting into the Bethlehem system was dispersed, now forming smaller blocks termed inclusions or enclaves. Enclaves resulting from interaction of two magma types give evidence of a heat source that could have melted metasedimentary rocks to form the initial Bethlehem magma.

17a. Kinsman Granodiorite, New Hampshire Plutonic Suite, Bradford, NH

Like the Bethlehem Granodiorite, the Kinsman Granodiorite plutons were also emplaced as thin sheets that were overridden by the Acadian Orogeny thrusts (Figure NHS-2). The Kinsman rocks were not as deformed as the Bethlehem Granodiorite; either the Kinsman magmas were emplaced later when thrusting was not as active, or the thrust fault into which the Bethlehem was emplaced was more active, lasting longer than that of the Kinsman. The Kinsman plutons are higher in the sequence, having overridden the Brennan Hill thrust and the Bethlehem plutons. There are locations in Kinsman Granodiorite plutons that were deformed after the magma completely solidified, but for the most part, the overlying thrust stopped moving while the magma was still partially molten. The alignment of elongated minerals in the Kinsman rocks occurred in the magmatic state.

The ages of the Bethlehem and Kinsman plutons are not defined with sufficient resolution to determine which is older. Each is about 410 million years old. The similar ages indicates that both thrust sheets were active at about the same time, perhaps with the Bethlehem sheet being slightly older than the Kinsman which could account for the longer duration of shearing of the Bethlehem.

Driving Directions

Take I-89 north to mile 20. Exit at the Warner-Bradford Exit and drive 7.6 miles on RT 103 to Bradford. Proceed through Bradford on RT 103 for 0.3 miles to large road cuts (Figure 17-1). Pull off to the right and examine cuts on the north side of the road (Figure 17-2).

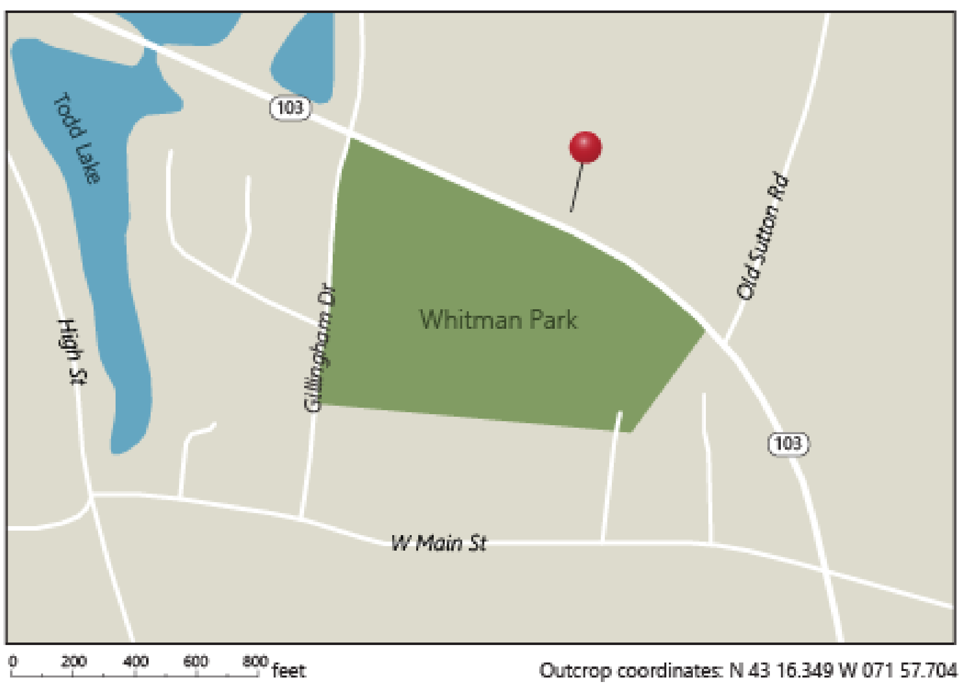


Figure 17-1. Map of the Bradford, NH area showing the location of the cuts for this stop.

Figure 17-2. General view of the cuts of the Cardigan Pluton at Bradford, NH (N43°16.349', W071°57.704').

On the Outcrop

The rocks of this road cut are in the Cardigan Pluton, one of several Kinsman Granodiorite plutons in New Hampshire (Location 17a of Figure NHS-1). Granodiorite is a granitic rock that contains more plagioclase feldspar than K feldspar (Figure I-5). The Kinsman Granodiorite is one of the most distinctive rocks in all of New England. It is easily recognized, even driving at highway speeds along I-89, from the large potassium feldspar phenocrysts. An excellent location exhibiting these feldspars is described below. This location is of interest because of the abundance of garnet in these igneous rocks.

Significance

A casual glance at this road cut shows some locations where the rock contains up to 80% garnet, such high amounts that geologists refer to the rock as garnetites rather than granodiorite (Figure 17-3A). A few meters away from the garnetite, the amount of garnet in the rock decreases but is still impressive (Figure 17-3B).

Garnet is better known as forming in metamorphic rocks that were originally clay-rich sediments, but is also found in granites of specific compositions. The magmas that formed the Cardigan Pluton of the Kinsman Granodiorite were derived by partial melting of metasedimentary rocks. The protolith (initial rock that was partially melted) was Al-rich from the abundance of clay in the original sediment, and these garnet-rich rocks represent the remains of the protoliths that were partially melted. The residue normally remains at depth as the magma escapes the source region as seen in the migmatites of Stop 15. This stop is unique in that the collision between Laurentia and Avalonia occurred with sufficient force to squeeze the magma and force it to ascend quickly, too fast for the denser garnet crystals to settle out from the magma and to be left behind at the site of melting. The magmas were squeezed up the Central New Hampshire Anticlinorium along with the thrust sheets of Central Maine Trough metasediments (Figure NHS-2).

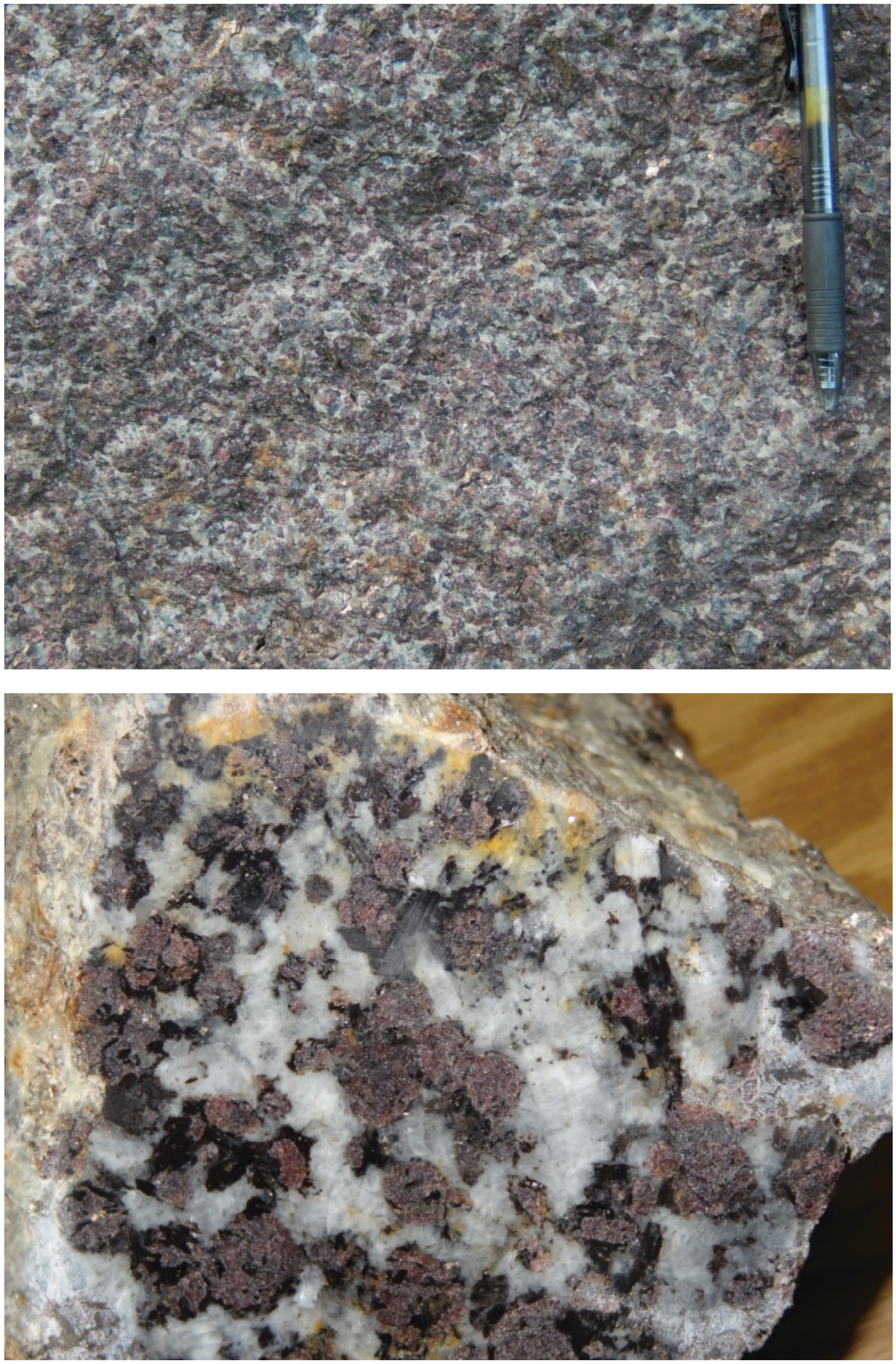


Figure 17-3A and 17-3B. The amount of garnet in the Kinsman Granodiorite is quite variable, from very high concentrations as show here to none at all in other locations.

17b. Kinsman Granodiorite, New Hampshire Plutonic Suite, Meredith, NH

Driving Directions

From I-93, take Exit 23 and follow RT 104 for 5.2 miles (Figure 17-4). Pull off on right and cautiously cross the road to cuts on the north.

Figure 17-4. Map of the cuts of the Meredith Granite near Meredith, NH.

On the Outcrop

The rocks at this location are part of the Meredith Porphyric Granite, a member of the Kinsman Granodiorite. Several Kinsman plutons are exposed from southwestern New Hampshire northward to Franconian Notch and beyond. As seen here, the Kinsman is easily recognized by the abundance of large, potassium feldspar crystals.

This road cut shows a spectacular concentration of elongated, white potassium feldspar crystals. Note that this particular road cut displays a very large concentration of feldspar crystals that are aligned in the same orientation (Figure 17-5). Walking a short distance from this location will take you to rocks that have lower feldspar abundances with less alignment. What makes this cut so interesting is that the feldspar crystals were concentrated by a process called filter pressing. The magma consisted of granitic liquid with abundant, suspended feldspar crystals. During crystallization, the number of crystals increased and the amount of liquid decreased to the point where the magma essentially acted as a semi-rigid framework. At that point, the Chesham Pond thrust overrode the pluton, smearing out the crystal-rich magma to the west. But the pluton wasn't fully crystallized like the Bethlehem Granodiorite of Stop 16. In the case of this road cut of the Kinsman, shearing of the partially ridged magma squeezed out granitic liquid from the framework, leaving a high concentration of oriented crystals behind.

Figure 17-5 (left). Abundant Kspar crystals in the Kinsman Granodiorite (N43º37.378', W071º 33.102').

Figure 17-6 (right). Carlsbad twin in a Ksapr crystal.

A close examination of the feldspar crystals shows that some of them are twinned (Figure 17-6). Twinned crystals have what appears to be essentially a mirror plane down the middle of the crystal where each half of the crystal is a mirror image of the other. Twinned Kspar crystals form during crystallization of magmas, not under metamorphic conditions. The twin plane boundary between the two sides of the crystal is not bent or broken, indicating that the crystal was not deformed to transpose it to this orientation. Note the difference between this crystal and the Kspar crystals in the previously described Bethlehem Granodiorite (Figure 16-3). Those crystals were deformed when the Bethlehem was completely solidified whereas all the feldspars of this outcrop were aligned when the pluton was still mushy, i.e., a mixture of crystals and liquid. Deformation aligned the crystals, expelled most of the liquid amongst the crystals, leaving behind rocks that are very concentrated in feldspar. This process occurred during thrusting of the overlying Chesham Pond thrust as it was pushed to the west during the Acadian Orogeny.

Another very interesting feature of the Kinsman Granodiorite is found a few tens of yards to the east. Here one sees a finer grained, dark rock that is intermingled with the Kinsman. Figure 17-7 shows fragments of the Kinsman engulfed in the darker rock and even individual Kspar crystals mixed in as well. This cut provides evidence that a more mafic magma mixed with the Kinsman magma. The Kspar crystals were originally in the Kinsman magma but upon mixing, were distributed in the mafic magma. Many of these isolated Kspar crystals are no longer rectangular, instead now have rounded or angular terminations. Rounding of the originally square ended crystals resulted from resorption of the crystals in the hotter, more mafic magma.

Figure 17-7. Magma mixing in the Kingsman Granodiorite (N43º37.364’, W071º33.007’).

This cut is interesting because it hints at a potential heat source that melted the metasedimentary source rocks that produced the Kinsman magmas. These dark rocks have compositions indicative of mantle-derived magmas. These magmas are hot enough to cause partial melting of the metasedimentary rocks they intrude.

Finally, this road cut shows dikes of Concord Granite cutting through the Kinsman (N43º37.364’, W071º32.998’). These younger granites lack aligned feldspars; the dike was emplaced after the thrusting to the west ceased.

We can therefore decipher a relative sequence of events at this road cut: 1) emplacement of a sheet of Kinsman Granodiorite magma that contained a large percentage of feldspar crystals; 2) Shearing of the magma from thrusting of the Chesham Pond thrust above the pluton. The thrust has been eroded away at this location, but can be seen farther to the west; 3) Filter pressing of the Kinsman magma, aligning the feldspars to the same orientation and, at the same time, squeezing much of the liquid out from the crystals, leaving a high concentration of crystals; 4) After thrusting ceased at the end of deformation of the Acadian Orogeny, Concord Granites were emplaced across New Hampshire. These granites are post-tectonic, meaning they were not emplaced during movement or deformation of the rocks they intruded. They therefore lack a fabric, instead have a random orientation of aligned crystals as seen in the Concord Granite dike seen here.

18. Spaulding Tonalite, New Hampshire Plutonic Suite, Pipers Cove, NH

Three members of the New Hampshire Plutonic Suite were emplaced as sheets of magma that were incorporated along the base of thrust sheets during the Acadian Orogeny. Geologists use the term syntectonic to describe the tectonic setting during the emplacement of these plutons, meaning the magma was emplaced during tectonic movement of the rocks that they intrude, causing syn-magmatic deformation of the igneous rocks. The Bethlehem Granodiorite of Stop 16 and the overriding Brennan Hill thrust was first emplaced farthest to the west. Next, the Kinsman Granodiorite (Stop 17) and the Chesham Pond thrust were emplaced over the Brennan Hill thrust. Finally, the magmas of the Spaulding Tonalite, the youngest of the three syntectonic mambers, and was emplaced along the Central New Hampshire Anticlinorium (Figure NHS-1). This zone represents the location from which the thrust sheets moved upward from deeper in the crust before being transposed subhorizonally to the west or east (Figure NHS-2).

Unlike the Bethlehem and Kinsman Granodiorites that were incorporated in thrust sheets to the west, this pluton of Spaulding Tonalite was transported with thrust sheets to the east. Other Spaulding plutons were thrust to the west, but all the Spaulding rocks are more closely emplaced near the Central New Hampshire Anticlinorium, and were not transported very far from where they ascended.

The Spaulding Tonalite is significant because it represents a change in magma composition from the earlier Bethlehem and Kinsman Granodiorites. The Spaulding Tonalite is the most mafic member of the New Hampshire Plutonic Suite. Tonalite is a granitic rock with considerably more plagioclase than K feldspar (Figure I-5). The Spaulding rocks were derived from melting of different source rocks than the Bethlehem and Kinsman Granodiorites.

Driving Directions

From Meredith, NH, follow RT 3 north for 6.6 miles to Pipers Cove (Figure 18-1). Park at pull off on right next to Squam Lake at N43°43.787', W071°33.463'. Use caution to cross the road to the cuts to the west (Figure 18-2) because southbound traffic can't see pedestrians until the cars round the curve.

Figure 18-1. Map of the Pipers Cove area along Squam Lake, NH.

Figure 18-2. General view of the road cuts at Pipers Cove (N43°43.814’, W071°33.971’).

On the Outcrop

Unlike the Bethlehem and Kinsman that contain abundant muscovite and garnet respectively, the Spaulding Tonalite is hornblende-bearing as is easily noted at this outcrop (Figure 18-3) and in the photomicrographs of Figure 18-4. The Spaulding Tonalite at this location is part of the Winnipesaukee Pluton. The pluton is compositionally zoned with hornblende-bearing rocks located along the western portions of the pluton. To the east, the amount of hornblende diminishes to where it is replaced by biotite. The most eastern portions of the pluton are more similar in composition to the Bethlehem and Kinsman rocks, containing both biotite and muscovite.

Figure 18-3 shows the abundance of amphibole in the Spaulding Tonalite at this location. Note the faint layering of the rock.

The Bethlehem and Kinsman magmas were derived from partial melting of metasediments. Although the eastern portion of the Winnipesaukee Pluton contain muscovite, the majority of this and other Spaulding Tonalite plutons were derived from more mafic source rocks, more likely representing partial melts of amphibolites, similar to the Ammonoosuc Volcanics of Stop 12. The amphibolitic source generated more mafic partial melts than other members of the New Hampshire Plutonic Suite, resulting in hornblende crystallization. Amphibolites are relatively poor in potassium, hence, melts derived from amphibolites tend to be Kspar-poor and plagioclase feldspar rich, which yielded the tonalite composition of these rocks.

An interesting feature of this outcrop is the subhorizonatal layering defined by coarser-grained layers separated by finer grained layers (Figure 18-3). Note that the hornblende crystals in the coarser-grained layers are not deformed but are equant in form. This suggests that the layering is not the result of metamorphic deformation but is igneous in origin. This layering is interpreted to result from shearing of a crystalline mush that contained a high amount of crystals in the magma. This shearing occurred when the overriding thrust sheet smeared the magma along its base, aligning the minerals to form crude layering. Motion of the overriding plate must have ceased before complete crystallization of the magma because no evidence of solid-state deformation like that seen in the Bethlehem Granodiorite is present here.

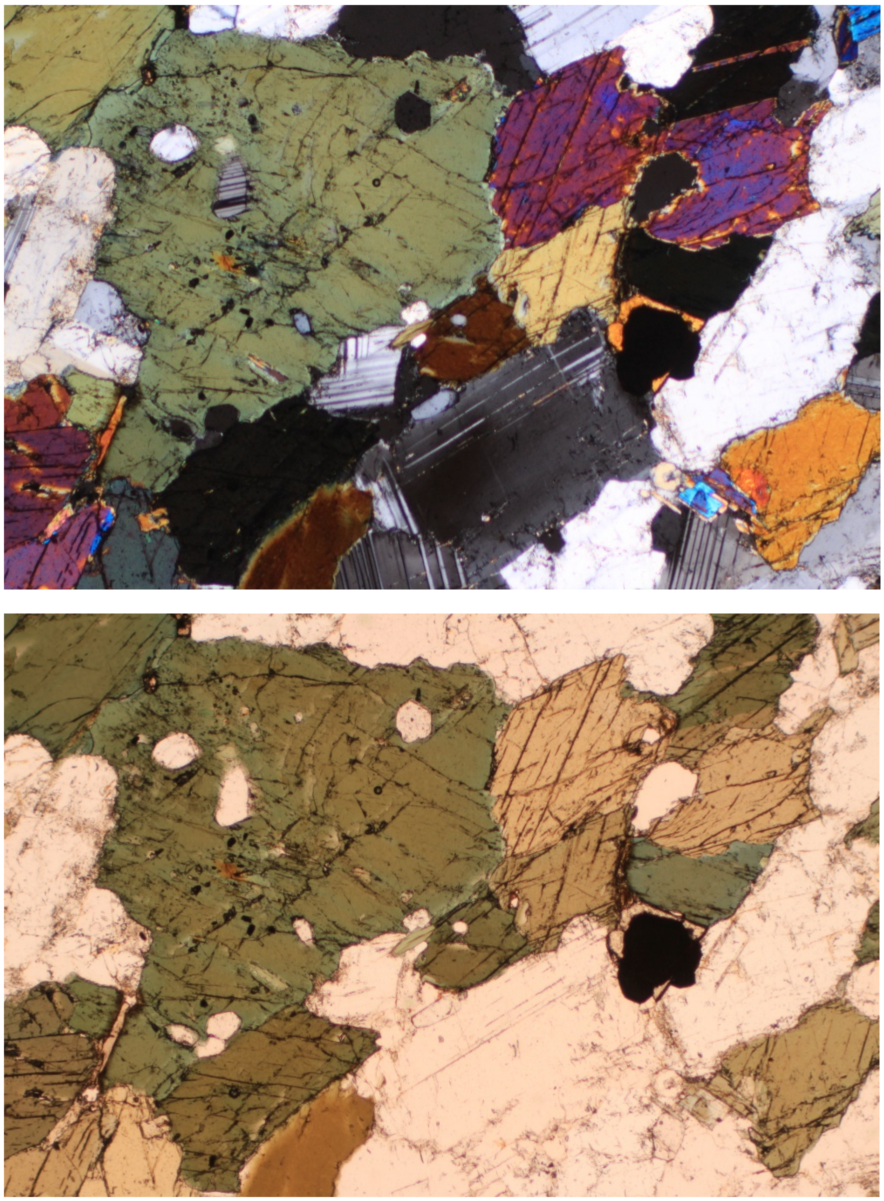


Figure 18-4. Crossed polarized (upper photo) and plane polarized light photomicrographs of the Spaulding Tonalite. In the upper photomicrograph, the green, purple and orange crystals are hornblende which is green and brown in the lower photo. Most of the gray colored mineral of the upper photo is plagioclase. Field of view is 5 mm.

19a. Concord Granite, New Hampshire Plutonic Suite, Bretton Woods, NH

The Concord Granite is the youngest member of the New Hampshire Plutonic Suite. Unlike the other three members that are syntectonic, i.e., the magmas were emplaced into rocks that were subjected to active tectonic processes, the Concord Granite intruded after tectonic motion ceased. Hence, these rocks lack the deformation of the Bethlehem Granodiorite and rarely show mineral alignment of the Kinsman and Spaulding members. They are classic representatives of two mica granites, containing both biotite and muscovite. The age of the Concord Granites ranges between 390 and 360 million years old.

Excellent outcrops of the Concord-type granite are found all across the state of New Hampshire. The most famous location to see the Concord Granite is at its type location in Concord. The Swenson Granite Works (369 North State Street, Concord, NH) has provided granite for decades. Older buildings in Concord and other NH towns commonly have Concord Granite as foundations. The quarried stone is also commonly used as curb stone along NH streets. So rather than giving in to the temptation to J-walk, look down and you might identify NH's most famous quarried rock. Liability concerns may prohibit visitors from entering the Swenson quarries, but superb, fresh granite is easily seen in the store yard.

My choice of the outcrops near Bretton Woods in northern New Hampshire was based on its beauty as well as being very representative of the Concord Granite. The Ammonoosuc River has carved a very scenic gorge that serves as a swimming hole for locals, known as the Upper Ammonoosuc Falls. Small cliffs of ~ 20 feet allow jumping into deep pools of crystal clear water. Bring a bathing suit and be prepared for cold water.

Most outcrops of Concord Granite are quite homogeneous, consisting of medium-grained feldspars, quartz, biotite and muscovite; one outcrop looks very much like another. Such is the case with the granite near Bretton Woods which is representative of the Concord Granite and one at a far less scenic location at a road cut at Hooksett, NH. The latter location, while lacking the beauty of Bretton Woods, has pegmatites containing 1 inch long tourmaline crystals and large books of muscovite.

Driving Directions

At Bretton Woods, take the Base Station Road towards the Cog Railroad station (Figure 19-1). At 2.2 miles up the road, pull off to the right at the poster board shown in Figure 19-2.

Figure 19-1. Map of the Bretton Woods, NH area with the red oval marking the location of the Concord Granite outcrops along the Ammonoosuc River.

Figure 19-2. General view of entrance to the foot bridge over the Ammonoosuc River.

Walking Directions

Follow the path to bridge over Ammonoosuc River. Cross the bridge (Figure 19-3), and descend to outcrops on the south side of the river to the left (N44º15.970’, W071º24.965’). During times of high water, use caution, some rocks along the river are slick.

Figure 19-3. General view of the gourge cut into the Concord Granite below the foot bridge.

On the Outcrop

In this rock, plagioclase feldspar is not easily distinguished from potassium feldspar (Figure 19-4). Both appear white. Quartz is the more translucent, gray mineral. The dark flakes are biotite, and characteristic of the Concord Granite is the silver-colored muscovite (Photomicrograph shown in Figure 19-5). Note the lack of alignment of the minerals that is easiest seen by the random alignment of biotite crystals. This randomness is indicative of post-tectonic plutons.

Figure 19-4. The characteristic mineralogy of the Concord-type Granites.

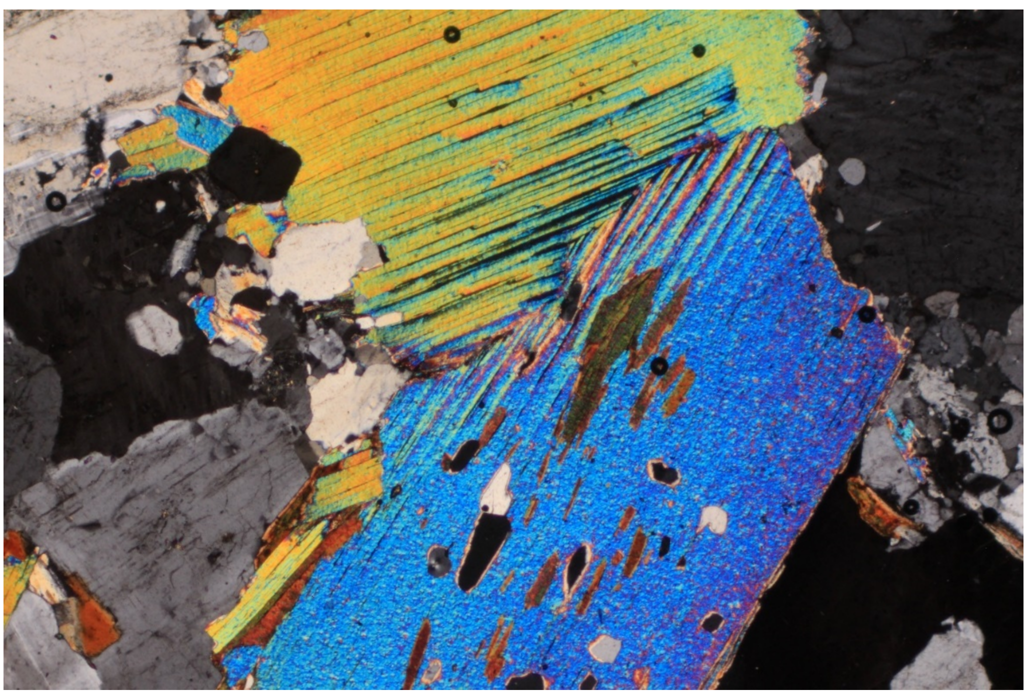


Figure 19-5. This crossed polarized light photomicrograph of the Concord Granite shows large crystals of muscovite with the orange and blue color. The blue muscovite contains darker, brownish green biotite. Field of view is 5 mm.

19b. Concord Granite, New Hampshire Plutonic Suite, Hooksett, NH

Driving Directions

From I-93 in south Concord, NH, take RT 3A exit and drive south for 9.5 miles (past the town of Hooksett). Turn right on road with sign Northeast Record Retention (opposite Sunrise Boulevard; Figure 19-6). The cuts of interest are on the north side of the road (Figure 19-7).

Figure 19-6. Map of the cuts of the Concord Granite, Hooksett, NH.

Figure 19-7. General view of the Concord Granite cuts (N43º03.500’, W071º27.827’).

On the Outcrop

This road cut has the same minerals seen in the rocks at the Upper Ammonoosuc Falls in Bretton Woods and other Concord Granite outcrops across the state. What makes this cut more interesting than others is the presence of pegmatites that cut the granite. Pegmatites are very coarse-grained rocks that represent late stage crystallization products of granitic plutons. They consist primarily of large quartz and feldspar crystals; at this cut some feldspar and quartz crystals are ~ 3 inches long. Particularly interesting are the large books of muscovite that are up to 2 inches long (Figure 19-8) and black tourmaline crystals that are up to 1 inch in length (Figure 19-9).

Tourmaline is a boron-bearing mineral. No previously crystallized minerals in the Concord Granite incorporated boron in their structures. These pegmatites represent the last traces of crystallization of the magma and are enriched in all elements excluded in minerals that formed earlier in the crystallization sequence, forming tourmaline and other interesting minerals in pegmatites in general. The neighboring state of Maine is famous for spectacular pegmatites, some with gem quality tourmaline and beryl, as is the Ruggles Mine in Grafton, NH and the Palermo Mine in Groton, NH.

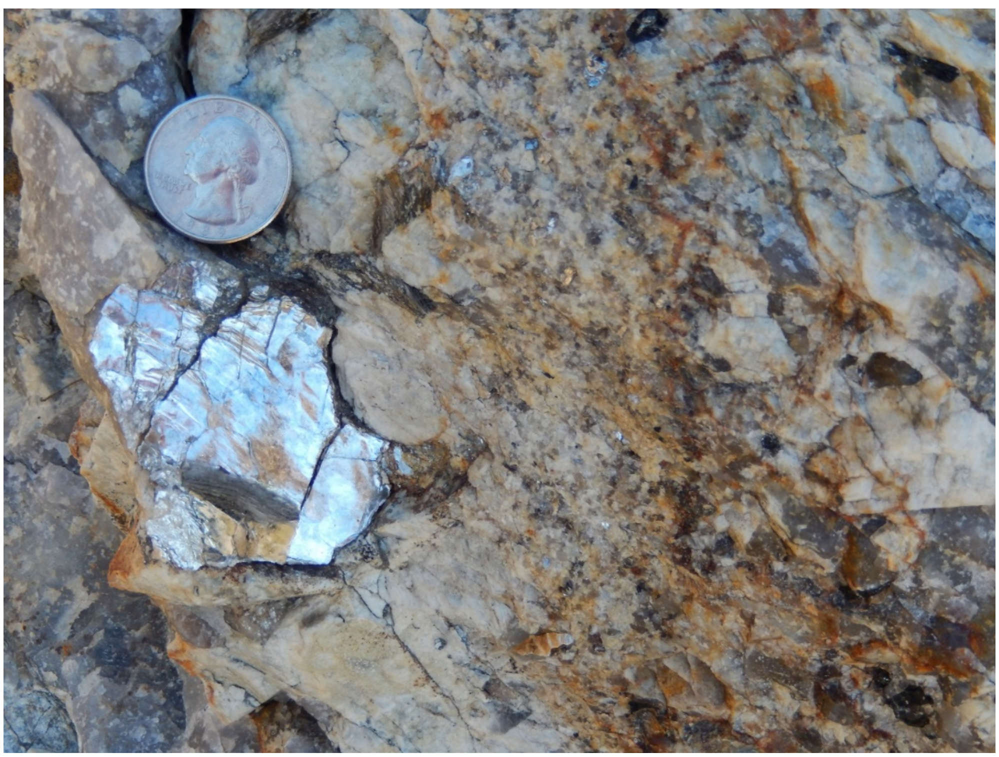


Figure 19-8. Photo showing a large muscovite in the pegmatite cutting the Concord Granite.

The presence of tourmaline requires source rocks that contained boron that were melted to form the granite. Boron-bearing sediments are deposited in oceanic settings, suggesting that the metasediments that melted to form the Concord magmas were deposited in an ocean setting along the margin of Laurentia. The presence of muscovite as an igneous mineral in these granites indicates that the magma was rich in aluminum, a characteristic of granites derived by partial melting of metamorphosed sedimentary rocks that were clay-rich.

Figure 19-9. Photo showing tourmaline crystals in the pegamite cutting the Concord Granite.

It is informative to compare the Concord Granite with older members of the New Hampshire Plutonic Suite. Stop 16 in the Bethlehem Granodiorite showed the strong fabric of the rock that was developed during the compressive events of the Acadian Orogeny (Figure 16-3). The Kinsman Granodiorite of Stop 17 also shows a fabric (Figure 17-5), but for the most part, it was developed during the magmatic state and not from solid-state deformation like that shown by the Bethlehem Granodiorite. The Spaulding Tonalite of Stop 18 shows even less of a fabric, indicating that it was emplaced during the tail end of Acadian deformation (Figure 18-4). In contrast, the Concord Granite at these stops lacks any deformation fabric. These plutons were emplaced after the Acadian compression ceased and are referred to as post-tectonic granites, i.e., they were emplaced after the tectonic motions of the Acadian Orogeny ceased.

20. Exeter Diorite, Merrimack Trough, Raymond, NH

The northern Appalachians experienced several collisional events as various island arcs, microcontinents, and eventually the Gondwanan continent collided with Laurentia. But before any of these tectonic elements could collide, the oceanic basins between them were consumed by subduction. As the oceanic plates subducted, under either another oceanic plate or under a continental margin, volcanic arcs were produced. The Acadian Orogeny imparted a strong metamorphic imprint across New Hampshire, but arc-related rocks resulting from subduction prior to the collision are not common in northern New England. Erosion has long removed any volcanic edifices that may have formed; now only the deeper seated plutonic roots of the arc remain. These occur as a series of plutons that extend from northeastern Massachusetts to southern Maine. The best example of these continental arc plutons in New Hampshire is the Exeter Diorite (Figure NHS-1).

New Hampshire is known as the Granite State after its abundant granitic plutons. Plutons dominated by more mafic rocks are not common, but the Exeter Diorite is the largest and best exposed mafic pluton of southeastern New Hampshire. Diorite is a plutonic rock with little to no Kspar and quartz and contains abundant plagioclase (Figure I-5). The pluton is named after its dominant rock type.

Good cuts of the Exeter Diorite are present at various locations on the University of New Hampshire campus and at road cuts adjacent to RT 101 farther to the south. The pluton shows a range in composition. Mafic rocks are most abundant, but the pluton ranges in composition from gabbro to more quartz- and Kspar-rich rocks that are present in the northern portions of the pluton.

Driving Directions

Follow RT 101 to Exit 9. Turn north on RT 27. At 0.1 miles, turn right on Watson Road. Proceed for 0.1 miles to road cuts on both sides of the road (Figure 20-1). Figure 20-2 shows the cut on the south side of the road.

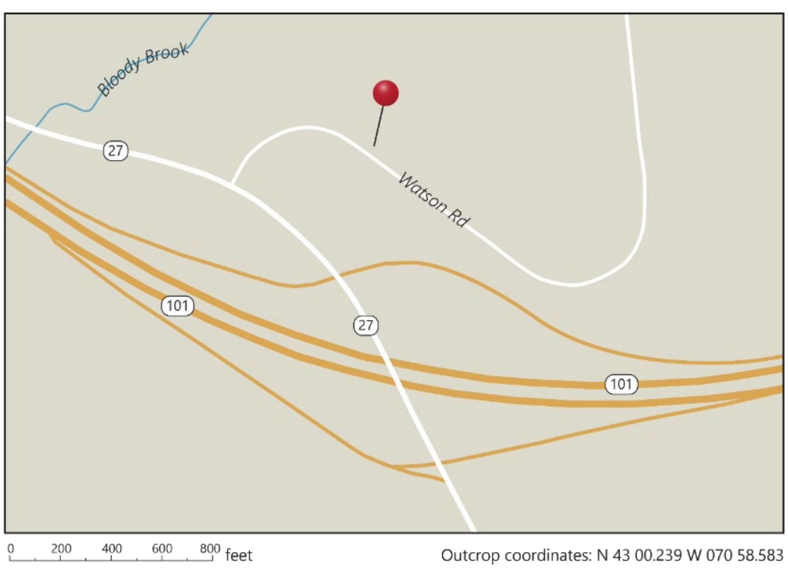


Figure 20-1. Map of the RT 101 Exit 9 area showing the cuts of the Exeter Diorite.

Figure 20-2. General view of the Exeter Diorite (N43º00.239', W070º58.583').

On the Outcrop

The rock at this cut are among the freshest of the pluton. Other locations, especially the northern portions of the pluton, have rocks that were altered by hydrothermal fluids expelled by the pluton immediately after crystallization of the magma. The fresh mafic rocks consist of orthopyroxene, clinopyroxene, hornblende, biotite, plagioclase and minor quartz. Figure 20-3 shows the fairly coarse-grain size of the rock. A photomicrograph shows that many of the pyroxene grains are rimmed by amphibole (Figure 20-4). Unlike pyroxene that lacks water in its structure, amphibole is a hydrous mineral, containing ~ 3 weight percent water. As the magma crystallized anhydrous minerals, the amount of water in the magma increased in the remaining magma. This caused a shift from pyroxene crystalizing to amphibole, accounting for amphibole rimming the earlier formed pyroxene.

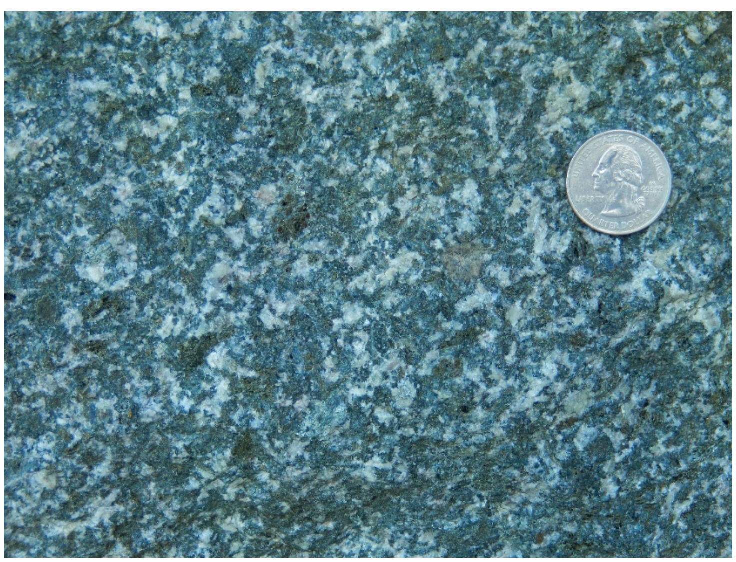


Figure 20-3. Typical rock of the Exeter Diorite.

One reason why the magma was hydrous is that oceanic plates experience metamorphism after the basaltic magmas are emplaced at mid oceanic ridges. Hot seawater alters the rock, forming a series of hydrous metamorphic minerals. As the oceanic plate subducted beneath the Laurentian margin just prior to the Acadian Orogeny, many of these hydrous metamorphic minerals exceed their temperature and pressure stability limits, causing a transformation to anhydrous minerals and releasing water that migrated upward. This water is incorporated into the magmas that form above the subducted plate, accounting for the higher water contents of the Exeter magma and eventually, to amphibole and biotite crystallizing instead of olivine and pyroxene.

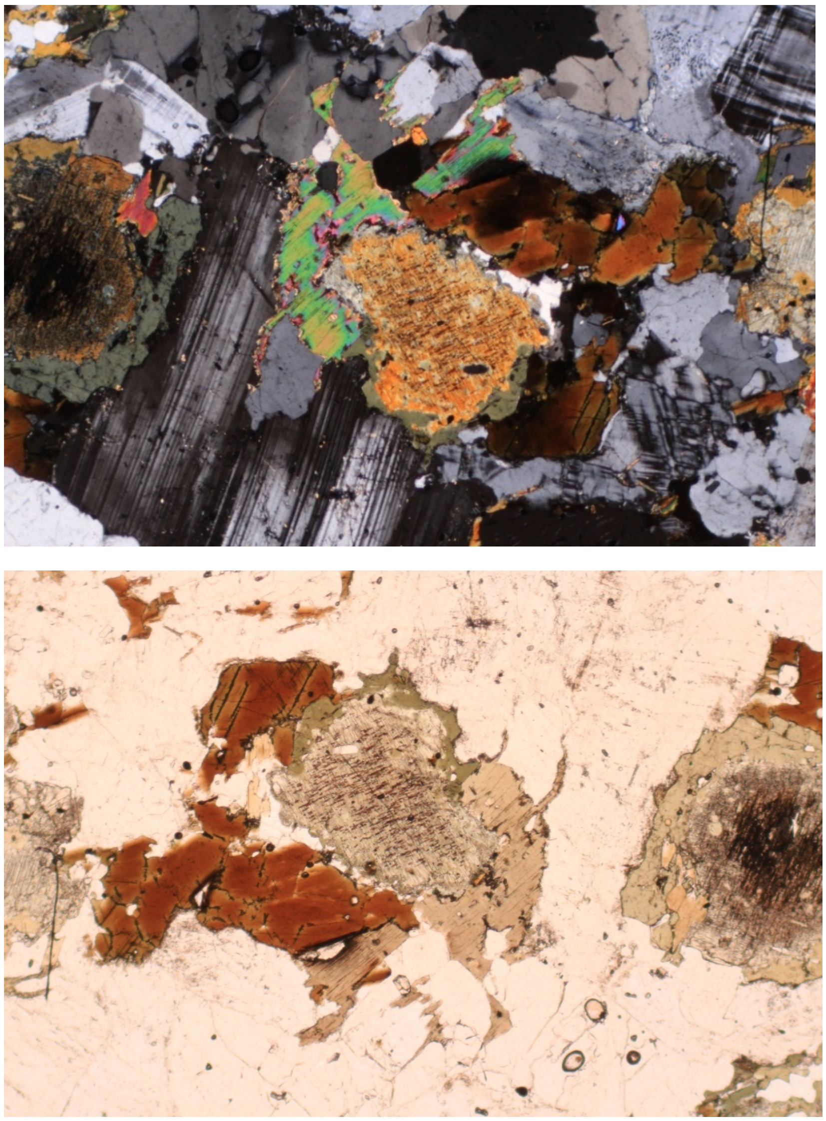


Figure 20-4. Upper photo is a crossed polarized light image of the Exeter Diorite, the lower photo is a plane polarized light image. The rock contains pyroxene (orange mineral in the middle of the photo), rimmed by amphibole (greenish colors). Other minerals include biotite (dark brown in lower photo) and plagioclase (striped black and gray mineral in upper photo). Field of view is 5 mm.

21a. Massabesic Gneiss Complex, New Hampshire’s Basement, Raymond, NH

The oldest rocks in New Hampshire are found in the Massabesic Gneiss Complex. While not as old as the Mount Holly Complex seen in Vermont at Stop 1, the Massabesic Gneiss Complex is also part of New England’s Precambrian basement. Geologists use the term basement to indicate the deepest known rocks that were subsequently covered by younger rocks, in this case, by the Paleozoic sediments of the Connecticut Valley trough (Stops 10 and 11) and the sediments of the Central Maine trough (Stops 14, 15, 27, 28, 29). But New Hampshire’s basement complex has a very different origin than the Precambrian basement rocks exposed in Vermont’s Green Mountains. The Green Mountains are cored by Precambrian rocks that represent the Laurentian crust, that is, those rocks were always part of the North American continent. In contrast, the Precambrian rocks of the Massabesic Gneiss Complex were formed on the opposite side of the Iapetus Ocean, adjacent to the Gondwanan continent. It is one of several fragments of Gondwana that rifted off that continent about 550 million years ago, much like Madagascar rifted off eastern Africa. The complex is part of a microcontinent called Ganderia that now forms the basement of coastal New England and portions of Maritime Canada. For a time span of 300 million years, the Iapetus and Rheic Oceans that separated Gondwana and Laurentia were consumed in several subduction zones, bringing the rifted piece of Ganderia that includes the Massabesic Gneiss Complex, to collide with Laurentia during the Salinic Orogeny. Sediments shed from the mountains raised from this collision, and from the remnant mountains from the earlier Taconic Orogeny, cover much of the accreted microcontinents in New England. Another microcontinental piece called Avalonia collided into Ganderia and the Laurentian margin to cause the Acadian Orogeny about 410 million years ago. Eventually, the continent of Gondwana collided with Laurentia and the previously accreted microcontinents to form the supercontinent Pangea about 290 million years ago.

Much of the complex shows large amounts of partial melting because it was deeply buried during the collision of Gondwana and heated to high temperatures, reaching granulite facies metamorphism (Figure I-6). The amount of partial melting in the complex varies; some portions appear to represent well over 50 percent melt, others far less. For that reason, two locations are describe here to give the reader a sense of the variable amounts of melting of the complex.

Driving Directions

Follow RT 101 to Raymond, NH at Exit 5. Turn north on RT 107. At 0.7 miles, turn left to follow RT 107 (Figure 21-1). Continue for 5.3 miles on RT 107 to large road cuts on the right (Figure 21-2).

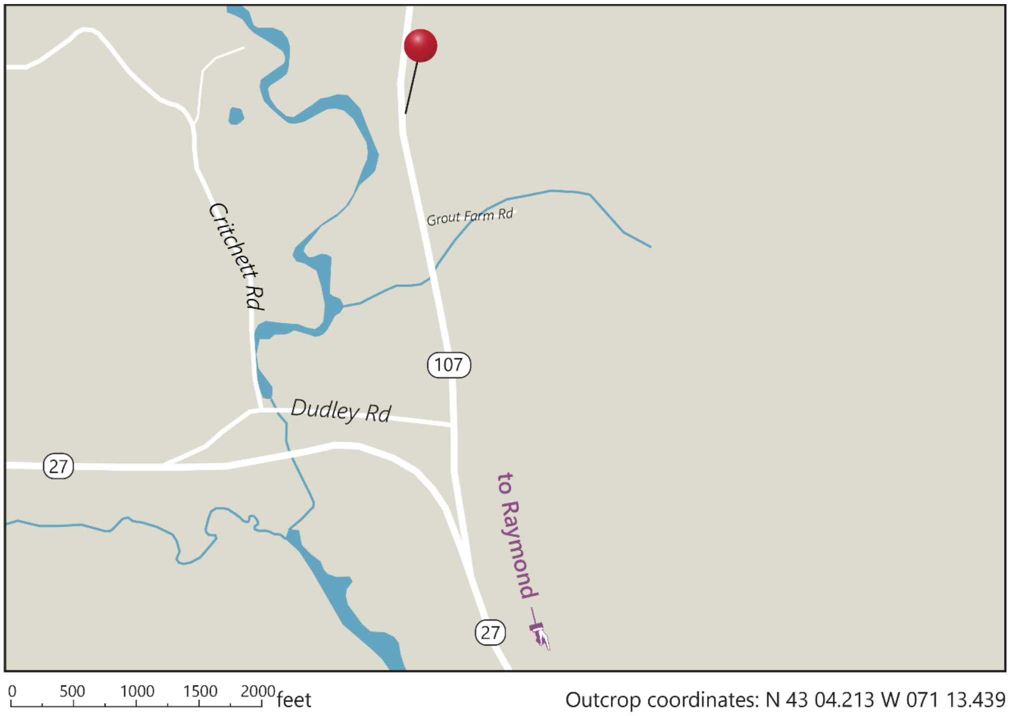


Figure 21-1. Map of the Raymond, NH area showing the road cut location to the northwest of Raymond.

Figure 21-2. General view of the Massabesic Gneiss at Raymond, NH (N43º04.213', W071º13.439').

At the Outcrop

This cut of the Massabesic Gneiss Complex is a migmatite with a very high percentage of leucosome (Figure 21-3). The leucosome represents the melted fraction of the rock, now seen as the light colored, feldspar- and quartz-rich bands. Unmelted, residual portions of the rock called melanosomes are the biotite-rich, darker swirly bands throughout the rock.

Geochronological studies revealed that the high grade melting event as seen here occurred during the Alleghanian Orogeny at about 290 Ma.

Figure 21-3. Photograph of the Massabesic Gneiss with abundant granitic leucosomes.

Both the leucosomes and the biotite-rich melanosomes were folded during the partial melting event. At high melt percentages, the rigidity of the migmatite was strongly reduced, allowing for the entire system to easily flow.

Given the high percentage of melting in this rock and the evidence that it occurred during the Alleghanian Orogeny about 290 Ma, one can question how we know anything about its prior history and the claims that it represents a fragment of a 600 million year old Precambriam basement. The complex contains zircon crystals that have three age populations. The oldest is 620 Ma, representing crystallization ages of magmas from the Precambrian island arc that formed adjacent to the Gondwanan continent. Erosion of the arc produced sediments containing a compositional signature of an arc source. These sediments were metamorphosed in the Acadian Orogeny and especially in the Alleghanian Orogeny when Pangea formed, forming additional zircon crystals with ages of ~400 and 290 Ma.

An additional feature of interest at this location is the Mesozoic dike that cuts upwardly through the crop (Figure 21-2). This dike is a precursor to the Onway Dike seen at Stop 22. It is a different composition than the Onway, having been emplaced several million years earlier during the very early rifting stage of Pangea, and is similar to the large dikes at Schoodic Point at Acadia National Park (Stop 35c).

21b. Massabesic Gneiss Complex, New Hampshire's Basement, Milford, NH

Driving Directions

From Milford, NH, follow RT 101A west. Just before the intersection with RT 101, park on the left in parking lot at the small shopping center (Figure 21-4).

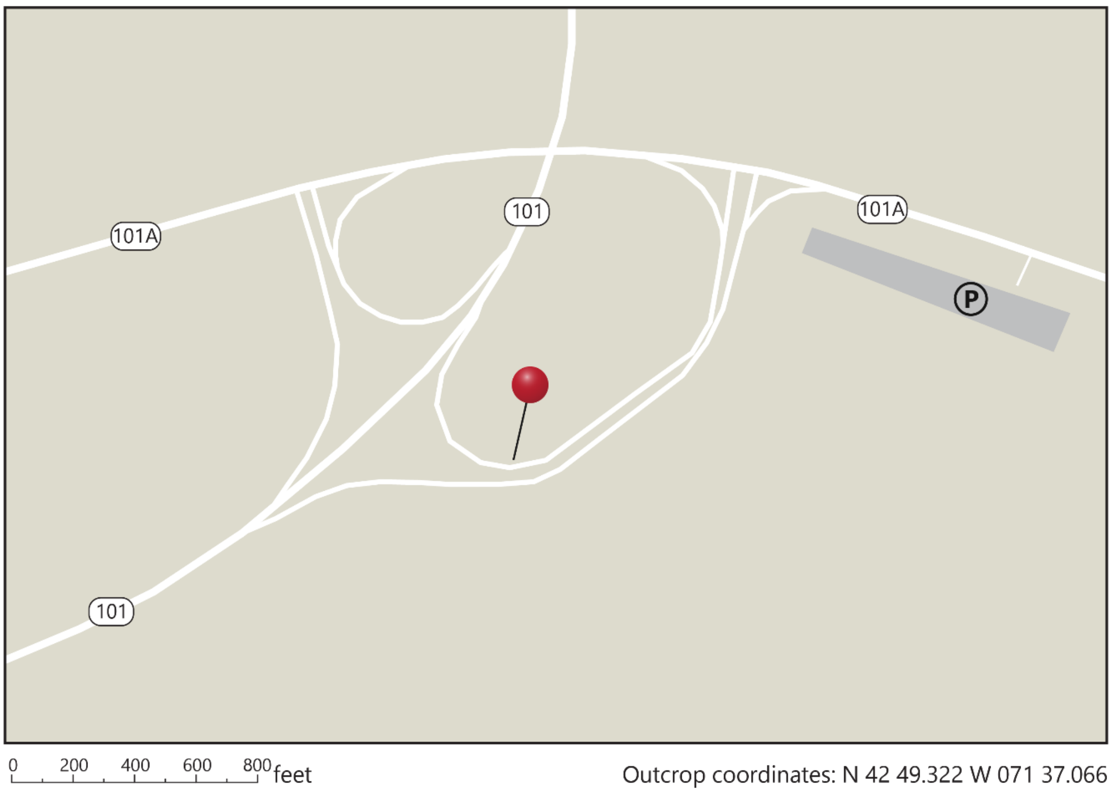


Figure 21-4. Map of the RT 101 and RT 101A junction showing the location of the cuts on the RT 101 onramp. Recommended parking location is shown by the red circle with the P.

Walking Directions

From the parking lot, walk west to the on ramp that leads to RT 101. Continue for 0.2 miles to where the on ramp diverges. Large road cuts are present along both sides of the road, but carefully cross to the north side of the road.

Figure 21-5. General view of road cut of Massabesic Gneiss (N42º49.322’, W071º37.066’).

At the Outcrop

These road cuts of the Massabesic Gneiss Complex are considerably different than the cuts of Stop 21a. The dark gray colored rock in Figure 21-5 is the same metamorphosed sediment seen earlier but note that the amount of light colored leucosome in the gray rock at this location is much lower than at Stop 21a, indicating that the amount of partial melting here was less. Figure 21-6 shows that the white leucosomes were pegmatitic, the large Kspar crystals have been disrupted, stretched and broken, revealing that here the rock was subjected to shearing and deformation after the leucosome completely crystallized. The feldspar crystals are no longer rectangular, but have thinned terminations from stretching.

But not all the rock at this cut was deformed. Dikes of undeformed granite cut the migmatite at the island to the southwest (N42º49.318’, W071º37.132’). This location still shows the tool marks where the granite was quarried decades ago. At many locations in this area, smaller dikes of this same granite cut the migmatite (Figure 21-7). This granite is the same as that in Milford and has been dated at 280 Ma, having been emplaced at the end of the Alleghanian Orogeny.

Figure 21-6. Deformed portion of the Massabesic Gneiss showing boudinaged (stretched and broken) pegmatitic leucosomes.

Figure 21-7. Undeformed, late-stage granitic dike cutting the Massabesic Gneiss.

Figure 21-8. Enclaves or inclusions of Massabesic Gneiss in late-stage pegmatite.

A dominant eye-catching aspect of this outcrop is the abundance of undeformed,younger pegmatite dikes that cut across the grain of the metamorphosed gray metasediments and the undeformed granite (Figure 21-5). Some locations show the metasediment as fragments that have been engulfed into the pegmatitic magma (Figure 21-8). But note that the pegmatite is not deformed and sheared like the pegmatitic leucosome of Figure 21-6.

These field relations yield the following history. First, the metasediments were partially melted and contained coarse-grained pegmatitic leucosomes that crystallized and were then deformed by shearing. No age has been determined for this event. At a later time, after the main compressive motion of Gondwana colliding with Laurentian ceased, the rock was intruded by the granitic sheets and then finally by the large, pegmatitic dikes that are undeformed and include fragments of the migmatite.

22. Onway Dike, Central Atlantic Magmatic Province, Onway, NH

The time was 201 million years ago, the setting, the supercontinent of Pangea. After a series of orogenic events along the eastern margin of Laurentia, the last collision occurred about 290 million years ago with the closure of the last remnant of the Rheic Ocean and the collision of Gondwana to the Laurentian continent to form the supercontinent Pangea (Figure O-13). Most of Earth's continental masses were joined at this time. But at 201 million years ago, one of the largest volcanic events in Earth's history occurred as Pangea began to rift apart to separate Pangea into Europe and Africa from North and South America as the Atlantic Ocean basin began to open (Figure O-14). This rifting thinned the lithospheric crust, allowing mantle upwelling and partial melting. Tremendous volumes of basaltic magmas ascended through major fractures as dikes to erupt as flood basalts called the Central Atlantic Magmatic Province (CAMP; Figure O-15). The flood basalt province was broken and dispersed as rifting carried the once united Pangea fragments into separate continents. Remnants of these massive flood basalts are found along the east coast of North America, in Brazil in South America and across the Atlantic in Spain in Europe and from Morocco to Liberia in western Africa (Figure O-15). In New England, most of the lava flows have been eroded away, though some are still preserved in the down-dropped Hartford Basin of Connecticut. But with this erosion comes a bonus, the dike system that feed the flows are revealed. The Onway Dike is one portion of a very long dike that feed the 201 Ma lava flows. This dike has several names, depending on location, but it is one long dike extending from Connecticut where it is called the Higganum Dike, through Massachusetts where it is called the Holden dike, to New Hampshire where it is named the Onway Dike after the nearby Onway Lake, to coastal Maine where it is named the Christmas Cove dike of Stop 42, and to Maritime Canada where erupted lava flows are spectacularly exposed at North Mountain in Nova Scotia (Stop 47 of Hild and Barr, 2015).

The voluminous eruption of basaltic magmas at 201 Ma may have caused the mass extinction event noted at the Triassic – Jurassic boundary. The event was one of the major extinction events of the Phanerozoic eon when over half the large number of life forms on Earth, both terrestrial and marine, disappeared. The extinctions left an ecological vacuum, allowing dinosaurs to evolve and become the dominant species for the next 150 million years. The cause of the extinctions may have been from massive amounts of CO_2, SO_2 and other climate influencing aerosols that were liberated into the atmosphere during the volcanic eruptions, causing world-wide climate change.

In 1815, Mount Tambora erupted in Indonesia, sending volcanic ash into the stratosphere. The amount of ash dispersed in the upper atmosphere was large enough to lower the temperature for months, even to the degree that 1816 is referred to as the year without a summer. Mount Tambora was a small volume eruption compared to CAMP. Tambora was also ash rich whereas CAMP was probably more gas-rich. Nonetheless, one can only imagine the atmospheric changes that resulted from thousands of cubic miles of basaltic lava erupting over a relatively short time period. New England may not be geologically very active now, being part of the passive margin of North America. But during the initial opening phases of the Atlantic Ocean basin, it was a very different story.

Driving Directions

Follow RT 101 to Exit 4. Turn north on Old Manchester Road. At 0.2 miles, turn left on Scribner Road for 0.4 miles, then turn right on Onway Road (Figure 22-1). At 0.1 miles, pull into the dirt road leading to the Gordon A. Cammett Jr Youth Recreation Area. If the gate is open, drive through and park at a small pull off to the right just past the gate. If the gate is closed, park along the dirt road.

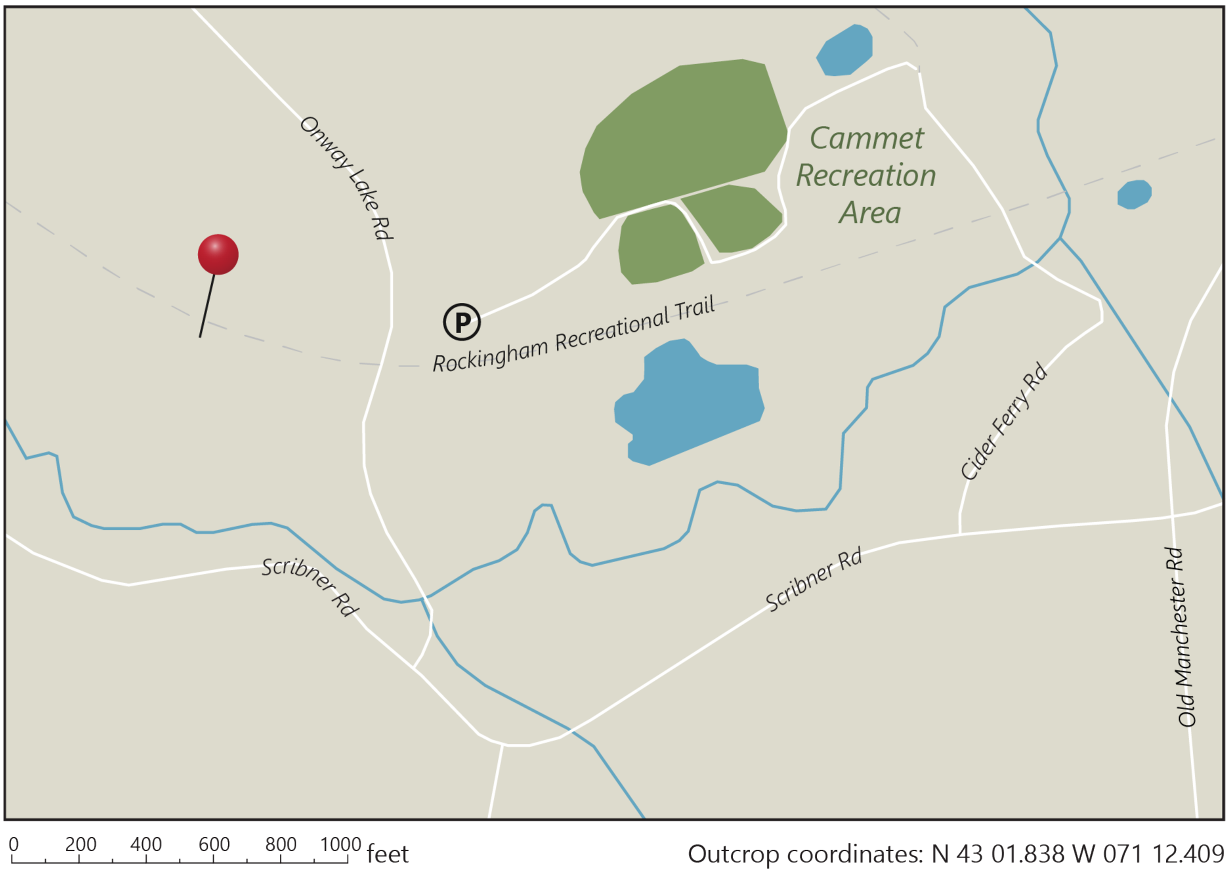


Figure 22-1. Map of the Onway Lake area, Onway, NH.

Walking Directions

If you parked in the small pull off past the gate, walk south to the abandoned railroad bed and turn right. Walk for ~200 feet to the bridge, passing under it and proceed for another 500 feet to the cuts (Figure 22-2). If the gate to the recreation area is closed, walk west across Onway Road and descend down to the railroad bed. The southern slope is easier to descent. You'll recognize the dike by its massive nature and the drill holes for paleomagnetic studies.

Figure 22-2. General view of the Onway Dike to the west of the bridge along the abandoned railroad track (N43º01.838’, W071º12.409’).

At the Outcrop

What this outcrop lacks in esthetic appeal because of its fine grain size, it more than compensates because of its geologic significance. The Onway Dike in this location is approximately 80 feet thick. A fracture that wide must have had massive volumes of basaltic magma pass through as thousands of cubic miles of basaltic lavas erupted through it to erupt on the surface. It is possible that the dike was originally wider because dikes tend to contract after the passage of magma. A similar CAMP dike in Maritime Canada averages 400 feet wide!

The effect of such large eruptions on the Earth’s climate must have been considerable, given that CO_2 is a greenhouse gas and SO_2 has the opposite, cooling effect. Massive amounts of both gasses were probably released during Central Atlantic Magmatic Province eruptions.

Figure 22-3 shows a photomicrograph of the rock from this outcrop. Abundant plagioclase, clinoporoxene, is shown by the yellow, orange, and blue colors. Orthopyroxene, a Ca-poor pyroxene, is the orange brown colored crystal with the dark fractures. Plagioclase is the gray to white elongated crystals.

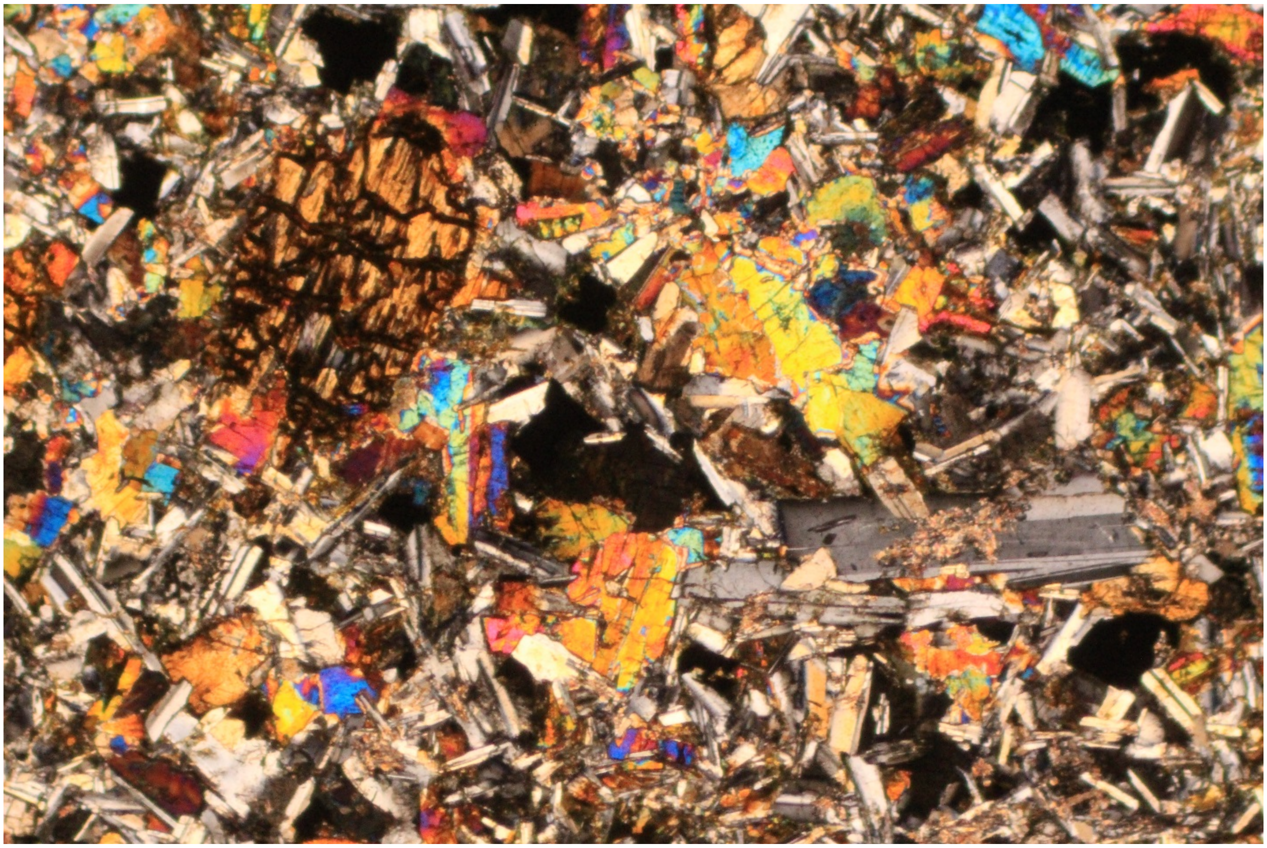


Figure 22-3. Phtomicrograph of a typical Onway Dike sample containing clinopyroxene, orthopyroxene, and plagioclase feldspar.

23. Albany Quartz Syenite, White Mountain Plutonic-Volcanic Suite, Jackson Falls, NH

After Pangea rifted and the Atlantic Ocean started to form, a Jurassic to Cretaceous pulse of magmatism occurred in linear belts as a chain of seamounts off the shore of New England, as plutons extending in a north, northwest alignment through New Hampshire, and in a west, northwesterly trend that define the Monteregian Hills across Quebec (Figure O-16). These rocks are not related to any mountain-building event, rather are the result of mantle-derived magmas ascending along a linear structural weakness. Many of these plutons were emplaced as ring dike complexes

New Hampshire is famous for its ring dike complexes. These circular dikes formed as a magma chamber ascended towards the surface. The overlying crust broke and subsided down into the chamber, producing major circular faults that mark the edges of the chamber. Magma ascended up the faults, forming the prominent ring dikes found in several complexes in New Hampshire (see Figure 26-2 for a schematic depiction of ring dike formation). These include the Belknap Mountains (Stop 25) and the Ossipee Mountain (Stop 26) Complexes as well as several other complexes in the White Mountains of New Hampshire (Figure 23-1). In some cases, the magma ascended though the ring faults to erupt on the surface as volcanoes. Examples of these volcanic rocks are seen at the Ossipee Mountain Complex of Stop 26. The outcrops at Jackson Falls are excellent examples of the magma that was emplace in the ring faults that now are preserved as circular dikes.

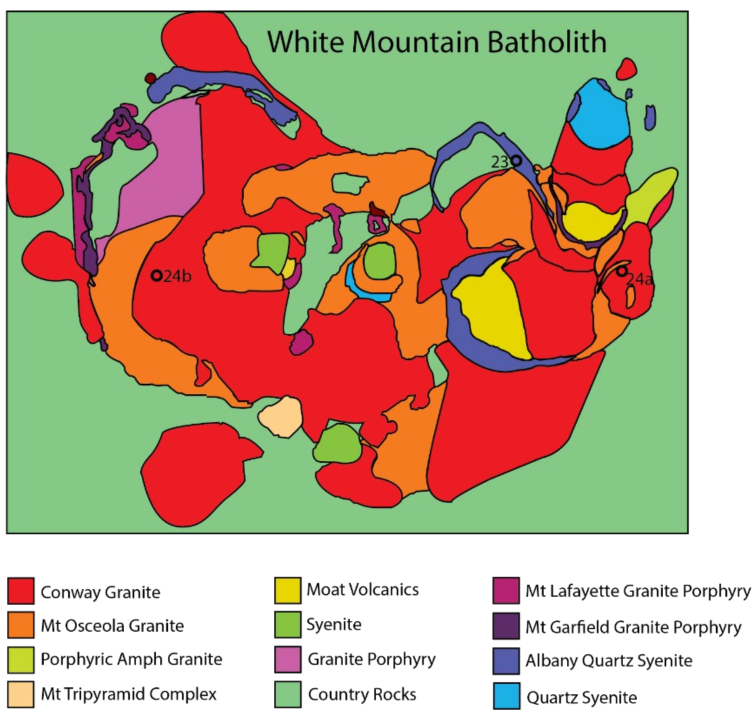


Figure 23-1. Generalized geologic map of the White Mountains Batholith (After Creasy and Fitzgerald, 1996). The eastern White Mountain batholith consists of three igneous centers, each with its respective ring dikes.

Driving Directions

From RT 16 at Jackson, NH, turn east and cross covered bridge to Jackson (Figure 23-2). Continue for 0.5 miles to Carter Notch Road and proceed 0.3 miles up Carter Notch Road to Jackson Falls on right. Park at the pull off on the right side of the road.

Figure 23-2. Map of the Jackson, NH area showing the location of the Albany Quartz Syenite outcrops at Jackson Falls.

Walking Directions

From the pull off, descent down to the river to the south (Figure 23-3).

Figure 23-3. General view of the outcrops at Jackson Falls (N44º09.046’, W071º10.807’).

On the Outcrop

Numerous interesting features are found in these outcrops. First is the Albany Quartz Syenite itself. While the layperson might be tempted to call this rock a granite, geologist would call these rocks quartz syenite. A syenite is a granite-like rock but it contains more Kspar than plagioclase feldspar and less than 5% quartz (Figure 23-4) whereas granites contain about the same amount of both feldspars with abundant quartz. The quartz syenite contains between 5 and 20% quartz, intermediate between a syenite and granite (Figure I-5). The distinction may seem trivial, but the differences indicate a very different origin for the Albany Quartz Syenite than many of the more quartz-rich granitic rocks across northern New England.

Figure 23-4. Photo of characteristic Albany Quartz Syenite.

Quartz syenites are common in the White Mountain Plutonic-Volcanic Suite. They are quartz-poor compared to granites because they do not represent melts of metasediments such as those found in the Rangeley Formation of Stop 15. Geologists think that these magmas were produced by mantle-derived mafic magmas intruding the crust, evolving to more silicic compositions and in some cases, mixing with crust melts as well. Evidence of a mafic magma component in the Albany Quartz Syenite is abundant as shown by the dark pillows of mafic rock throughout this outcrop (Figure 23-5). These dark rocks, called mafic inclusions or mafic enclaves, are thought to represent more mafic magma that was injected into the Albany Quartz Syenite when the latter was still molted. Because the mafic magma was at a higher temperature than the Albany, it partially quenched to a finer grain size. In some cases, the mafic magma was still somewhat molten and during emplacement, incorporated feldspars of the Albany. Many of the mafic inclusions at this location contain Kspar crystals that didn't crystallize in the mafic magma itself, but were stirred in from the quartz syenite.

Figure 23-5. Magmatic enclaves in the Albany Quartz Syenite. The enclave in the lower right portion of the photo contains Kspar crystals that were mixed into the more mafic magma.

The mafic enclaves are not the only fragments found in the Albany Quartz Syenite. Careful examination will reveal fragments of layered rock. These are broken portions of the rock into which the Albany was emplaced. Down dropping of rocks above the magma chamber created the circular fault that was intruded by the Albany. Many broken blocks resided along the fault, some of which floated in the Albany magma (Figure 23-6).

Figure 23-6. Fragment or xenolith of wall rock in the Albany Quartz Sytenite. N44º09.046’, W071º10.807’

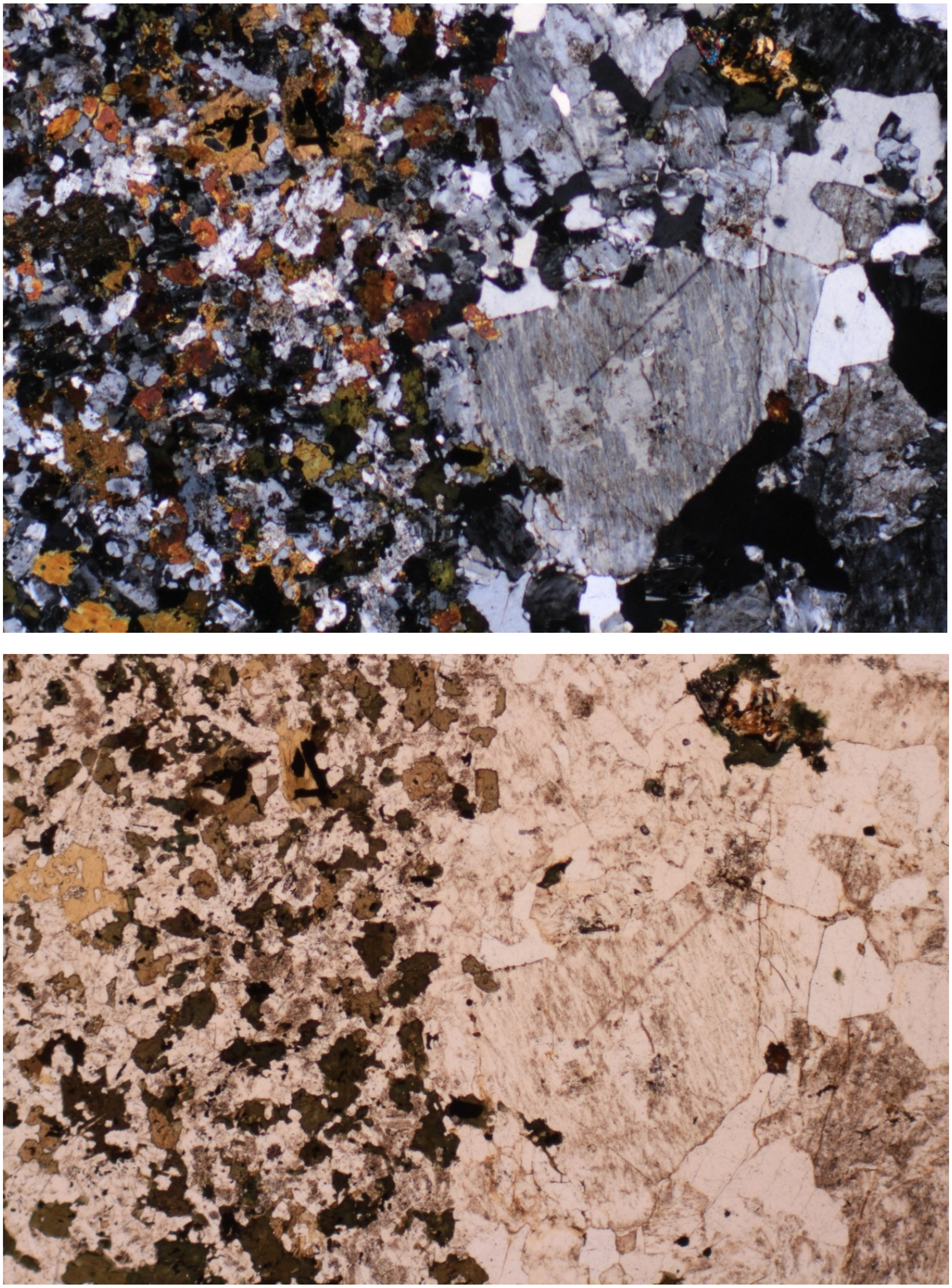


Figure 23-7. Crossed polarized (upper photo) and plane light (lower photo) photomicrographs of the Albany Quartz Syenite (right side of photos) and a finer grained, mafic enclave (left side of photos). The large, gray mineral is Kspar. The enclave contains abundant amphibole. Field of view is 5 mm.

24a. Conway Granite, White Mountain Plutonic – Volcanic Suite, Redstone Quarry, Redstone, NH

In 1886, the North Jay Granite Company bought land on Rattlesnake Mountain and developed the Redstone Quarry. Transporting large volumes of quarried stone required a railroad spur that connected to the nearby Maine Central line and the stone was then shipped all across the United States. Redstone Granite was used in many buildings in nearby New England in Portland and Boston, and at farther locations in New York, Washington, D.C., Denver, CO and even in Havana, Cuba. Some of the famous buildings and monuments constructed from the Redstone Granite include Grant's Tomb in New York (Yes, we know who is buried there – but technically, no one is buried in Grant's tomb because Grant and his wife are entombed in sarcophagi above ground rather than being buried in the ground) and the National Archives building in Washington

The Conway Granite is a classic example of what geologist term an "anorogenic" granite. This term means that the granite wasn't produced from collisional plate tectonic processes like the New Hampshire Plutonic Suite (Stops 16-19) or the Exeter Diorite (Stop 20). Rather, anorogenic granites are formed in settings unrelated to orogenic processes, i.e., in a non-mountain building setting, typically in a rifting or extensional environment. These types of granite typically require a mafic heat source to either partially melt lower crustal rocks that have previously been melted, already having lost a more typical granitic component, or from evolution of alkalic basalts to high silica residual magmas, usually combined with assimilation of crustal materials. Some of the White Mountain Plutonic – Volcanic Suite complexes, such as the Belknap Mountains and the Ossipee complexes, have Conway Granite associated with mafic rocks, suggesting that mafic magmas had a role in the origin of the Conway. But most of the White Mountain batholith lacks abundant mafic rocks, in fact they are quite rare. Any role of mafic magmas in the generation of the Conway here is speculative.

The Conway Granite is found at several White Mountain Volcanic and Plutonic Suite complexes (Figure 23-1), as far south as the Pawtuckaway Mountains in southeastern New Hampshire, and as far north as the Percy Peaks near Stark, New Hampshire. The north-northwest trend of the White Mountain Volcanic and Plutonic Suite complexes suggest that the magmas were intruded along a zone of major structural weakness.

The age and grain size of the Conway Granite varies a bit from location to location, but this quarry consists of classic Conway that is instantly recognized by anyone who has seen the rock elsewhere. The other abundant granite of the White Mountain batholith is the Mt. Oceola Granite (Figure 23-1). The Conway and Mt. Oscoela Granites differ in that the Conway contains biotite whereas the Mt Osceola has amphibole and biotite. Both rock types are present in the Redstone Quarry.

Driving Directions

From the town of North Conway, drive south on RT 16. Turn east (right) on Eastman Road (also marked as RT 302), and continue for 0.6 miles to Mountain Road on left (Figure 24-1). Follow Mountain Road for 0.2 miles to the end and park (Figure 24-2).

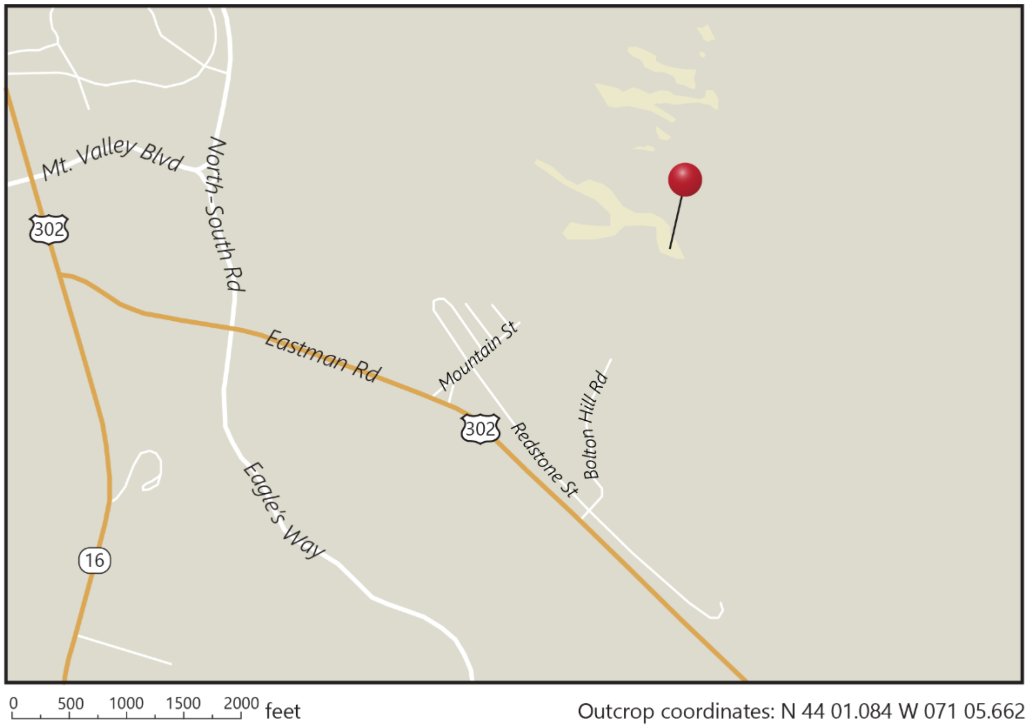


Figure 24-1. Map of the Redstone, NH area showing the location of the Redstone Quarry.

Figure 24-2. Sign at the end of Mountain Road at parking area.

Walking Directions

From the end of Mountain Road, follow the gravel road for 0.17 miles to N44º01.029’, W071º05.786’. A trail cuts across the road at this point, follow the trail uphill for a couple hundred yards to the quarry (Figure 24-3). Some scrambling over and around granitic blocks is necessary.

Figure 24-3. General view of Conway Granite at the quarry (N44º01.084’, W071º05.662’).

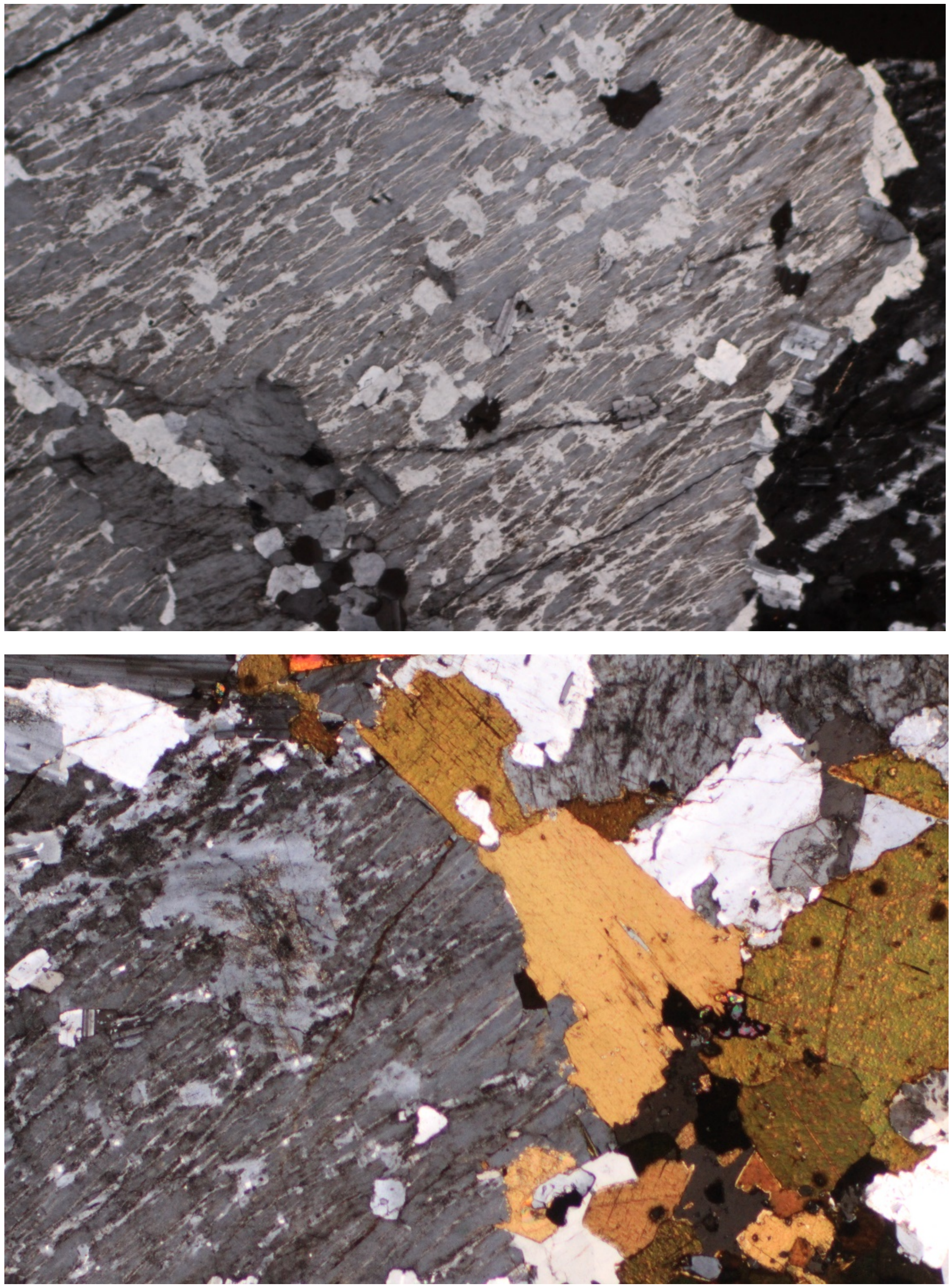
At the Outcrop

A feature of the Conway Granite that immediately catches one’s eye is the coarse-grain size (Figure 24-4). Unlike the Concord Granite, the other famous quarried New Hampshire granite, the Conway Granite lacks muscovite and is biotite-bearing. It also differs in that plagioclase feldspar is not abundant; the dominant feldspar is Kspar. Some of the Kpar crystals in the Conway are pinkish in color. The color is the result of hydrothermal fluids in the crystallized but still hot rock shortly after magma solidification that altered the small amounts of iron in the Kspar structure. The iron was oxidized to microscopic hematite crystals, imparting the pinkish color to the crystals. Some feldspar crystals show internal zoning that was produced as each successive layer grew on the crystal. 1/3 inch-size quartz crystals are also abundant.

Figure 24-4. View of typical Conway Granite at the Redstone Quarry.

Walk NW for 0.25 miles to N44º01.020, W071º05.872' to site with abandoned derricks to see the greenish variety of granite. This is the Mt. Osceola Granite that contains amphibole.

The photomicrographs of Figures 24-5A and B show an interesting feature of Kspar. At the high temperatures of crystallization from a magma, Kspar incorporates small amounts of plagioclase in its structure. But upon cooling, the plagioclase no longer fits in the Kspar structure and exsolves as thin bands of plagioclase feldspar, called lamellae, in the Kspar host crystal. In these photos, the Kspar is dark gray, the plagioclase lamellae are lighter gray. Sometimes, as is the case here, the exsolved plagioclase forms patches. Kspar with plagioclase lamellae is called perthite. The brownish to greenish brown mineral to the right of the lower photo is biotite.



Figures 24-5A and 24-5B. Crossed polarized light photomicrographs of Conway Granite showing pertitic potassium feldspar with albitic lamellae (5A) and biotite (5B). Field of view is 5 mm.

24b. Conway Granite, White Mountain Plutonic – Volcanic Suite, The Basin, Franconia Notch, NH

Driving Directions

Follow I-93 north to Franconia Notch State Park. Exit at the Basin at Highway mile 106.4. If travelling from the north, exit I-93 at mile 107 (Figure 24-6).

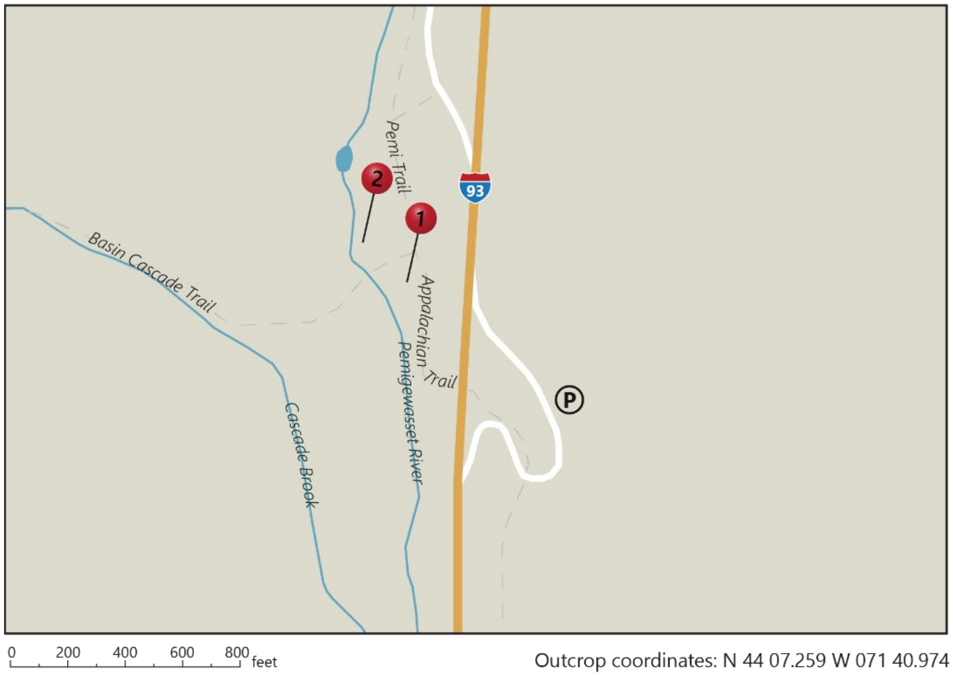


Figure 24-6. Map of the Basin, Franconia Notch, NH

Walking Directions

For those who parked on the east parking lot, walk through the tunnel under I-93. Follow the signs to the Basin.

On the Outcrop

Conway Granite is seen at the Basin viewing area (Figure 24-7). The rock at the Basin is very similar to the Conway Granite at the Redstone Quarry.

An interesting intrusion breccia consisting of Conway Granite is located north of the Basin (Figure 24-8). Along the margin of this Conway Granite pluton are many fragments of broken Kinsman Granodiorite. The blocks are angular, some are like puzzle pieces that would fit together if the interlaying Conway Granite were removed. One mechanism for creating space for granitic magmas to intrude the crust and assemble to form a pluton is by stoping. Stoping is a process whereby the country rocks are fragmented, forming blocks or xenoliths as seen here. As the xenoliths sink into the magma chamber, the magma migrates upward to take the place of the stoped block.

Figure 24-7. View of the Basin, a large pot hole carved by glacial waters into the Conway Granite (N44º 07.259’, W071º 40.974’).

Figure 24-8. Intrusion breccia in the Conway Granite (N44º 07.297’, W071º 41.006’).

25. Belknap Syenite, White Mountain Plutonic – Volcanic Suite, Alton Bay, NH

Another excellent example of a White Mountain Volcanic and Plutonic Suite complex is the Belknap Mountains Complex along the south shore of Lake Winnipesaukee. While lacking the completely circular ring dike of the Ossipee Complex, the Belknap Mountains Complex has more easily visited exposures in abundant road cuts along RT 11. For those with access to a boat, visiting Diamond and Rattlesnake Islands at the southern portion of the lake is enjoyable; these islands are segments of a ring dike composed of the Lake Quartz Syenite (Figure 25-2).

This road cut at this stop is interesting because it not only shows rocks of the White Mountain Plutonic – Volcanic Suite, but it shows their method of emplacement as well. The Belknap Syenite seen here is clearly a dike, cutting older pegmatitic dike of the New Hampshire Plutonic Suite. This cut offers a bonus because it also contains a much younger, Mesozoic mafic dike. These dark Mesozoic dikes are present at several road cuts along RT 11. They represent the initial mantle-derived magmas resulting from the rifting that split Pangea and led to the formation and growth of the Atlantic Ocean basin. Similar dikes are seen at Stops 21a, 35c and 38. So this road cut shows a dike cutting a dike cutting a dike. Dikes cubed!

Driving Directions

From the town of Alton Bay, NH, drive north for 5.4 miles to road cut on northeast side of road (Figure 25-1). From the opposite direction, follow RT 11 past the Laconia Airport through West Alton. A couple hundred yards past the RT 11 and RT 11D junction (on left) is a road cut (Figure 25-3).

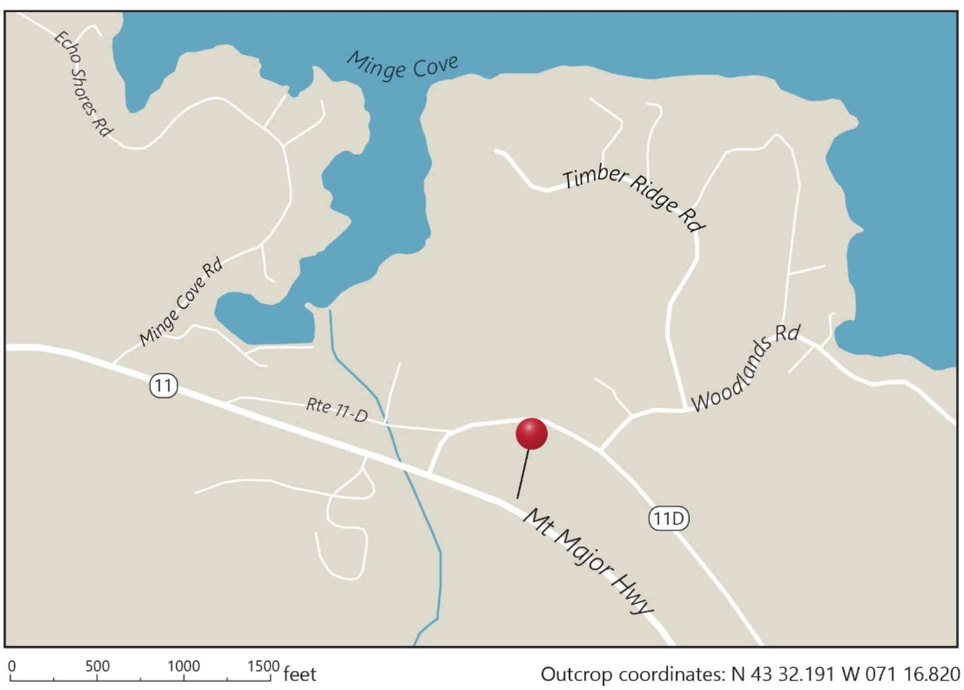


Figure 25-1. Map of the road cuts of interest along RT 11, north of Alton Bay, NH.

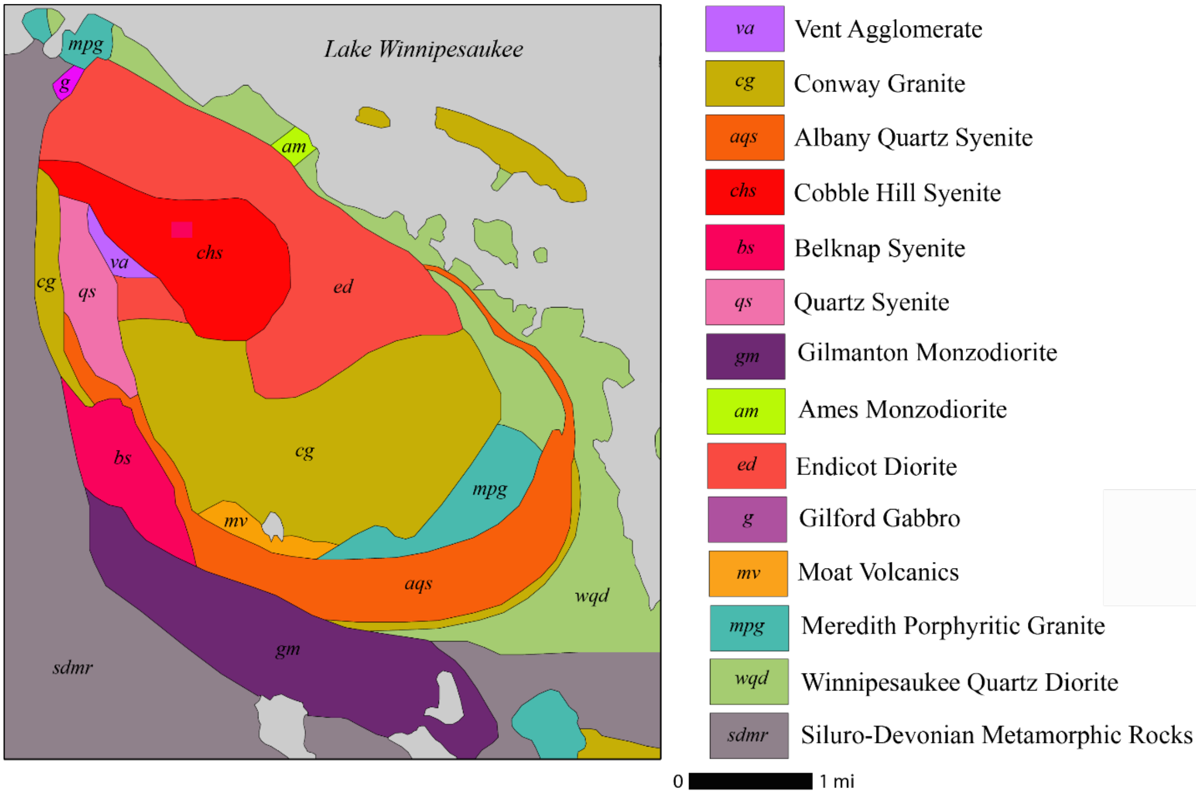


Figure 25-2. Geologic map of the Belknap Mountains Complex (After Bothner and Loiselle, 1987).

Figure 25-3. Road cut exposing the Belknap Syenite to the left with its characteristic brownish color. Immediately to the right of the Belknap Syenite are white colored rocks of the older, New Hampshire Plutonic Suite. This appears to be a pegmatite dike. The youngest dike is the black colored, Mesozoic dike that diagonals to the upper right (N43°32.191', W071°16.820').

On the Outcrop

The Belknap Syenite (Figure 25-4), a rock that is very quartz-poor (Figure I-5), shows interesting textural variations. Along the northern portion of the dike at the top of the cut, the grain size is finer than the dike's interior. The finer grain size is the result of chilling of the Belknap magma against the relatively cold New Hampshire Plutonic Suite rocks. Additionally, the margins of the dike appear sheared from movement of magma in the interior of the dike. The coarser-grained minerals in the dike's interior results from continued movement of the magma in the interior of the dike while the margins are more viscous, having been partially quenched. Drag of more viscous magma along the margins of the dike set a shear gradient; any crystals in the margin tend to migrate to zones of less shear stress, essentially tumbling out of the sheared margins to the interior of the dike.

The Belknap Syenite is similar to the Albany Quartz Syenite seen at Stop 23 in that it contains abundant Kspar as the dominant feldspar. While the difference may seem trivial to the non-geologist, the rock here is called a syenite versus the previously seen quartz syenite of Stop 23. The distinction is based on the amount of quartz present in the rock. Syenites contain less than 5 % quartz, quartz syenites contain between 5 and 20 %; quartz contents greater than 20% are granites (Figure I-5). Just as baseball fans distinguish shortstops from first basemen rather than lump them as infielders, so does the distinction of the granite-looking rocks have significance. Unlike the Concord Granite seen at Stop 19 that formed from partial melting of metasediments, syenites and quartz syenites have a different parentage. These quartz-poor magmas formed from mantle-derived basalts, being the end result of the magma crystallizing a long history of minerals to produce a Kspar-rich magma.

Figure 25-4. Photo of characteristic Belknap Syenite.

Figure 25-5. Faint grain size variations are evident in this rock with slightly darker and finer grained rock in the lower right portion of the photo.

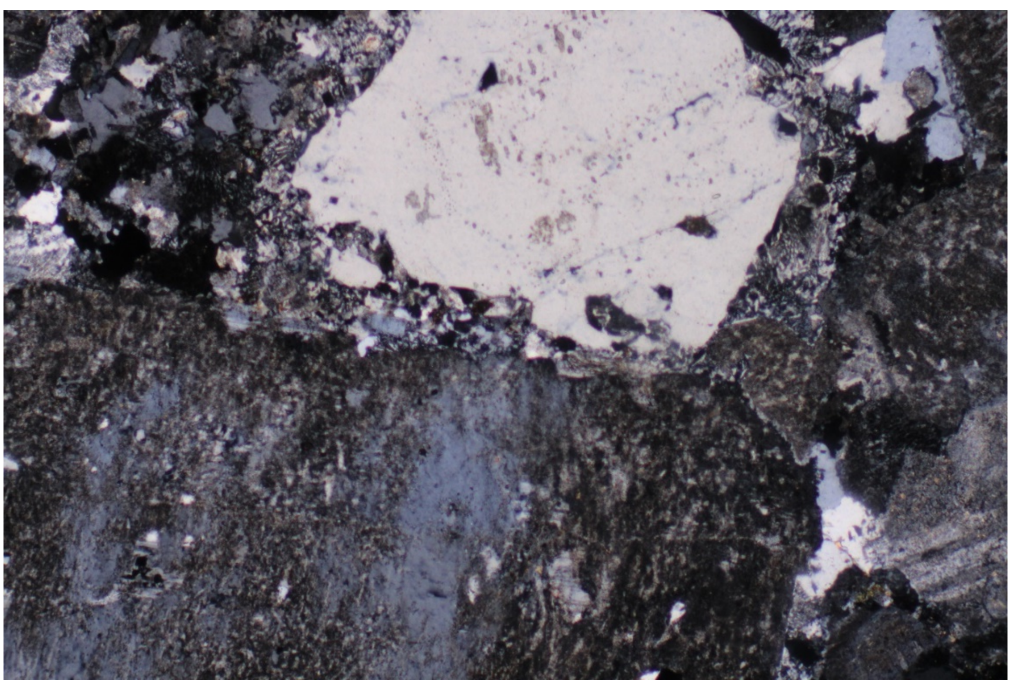


Figure 25-6. Crossed polarized light photomicrograph of the Belknap Syenite. The large gray colored crystal along the bottom of the photo is Kspar. A clear quartz crystal occupies the top center of the photo. A fine-grained matrix of feldspar and quartz surrounds the larger quartz crystal. Field of view is 5 mm.

Figure 25-7. Pegmatitic dike of the New Hampshire Plutonic Suite.

The evidence of mafic magma interaction with the Belknap Syenite is not as evident at this cut as it is in the Albany Quartz Syenite, but it is present. Look for subtle differences in the amount of dark minerals of finer-grain size than the typical Belknap Syenite (Figure 25-5).

On each side of the Belknap Syenite dike, one can examine rocks of the New Hampshire Plutonic Suite. Beautiful pegmatitic rocks (Figure 25-7) with white feldspar and quartz that forms a graphic granite in some locations (N43º32.187', W071º16.822'). Graphic granite is a texture that resembles cuneiform writing. It appears that the New Hampshire Plutonic Suite pegmatitic dike served as a zone of weakness, allowing the Belknap Syenite magma to exploit the weakness to form its own dike. Later, the Mesozoic diabase dike intruded close to this same zone of weakness.

26. Ossipee Ring Complex, White Mountain Plutonic – Volcanic Suite, South Tamworth, NH

After Pangea rifted and the Atlantic Ocean started to form, a Jurassic to Cretaceous pulse of magmatism occurred in linear belts, forming a chain of seamounts off the shore of New England, as plutons extending in a north, northwest alignment through New Hampshire, and in a west, northwesterly trend that define the Monteregian Hills plutons across Quebec (Figure O-16). These rocks are not related to any mountain-building event, rather are the result of mantle-derived magmas ascending along a linear structural weakness. Many of the intrusions form ring dike complexes.

New Hampshire is famous for its ring dike complexes and none is as famous as the Ossipee Ring Dike Complex (Figure 26-1). This complex was prominently featured in mid 20th Century controversies over the method of pluton emplacement, that is, how granitic magmas are emplaced into the crust to form plutons. One of the more famous debates in the science of geology is how is space created for such large volumes of granitic magma.

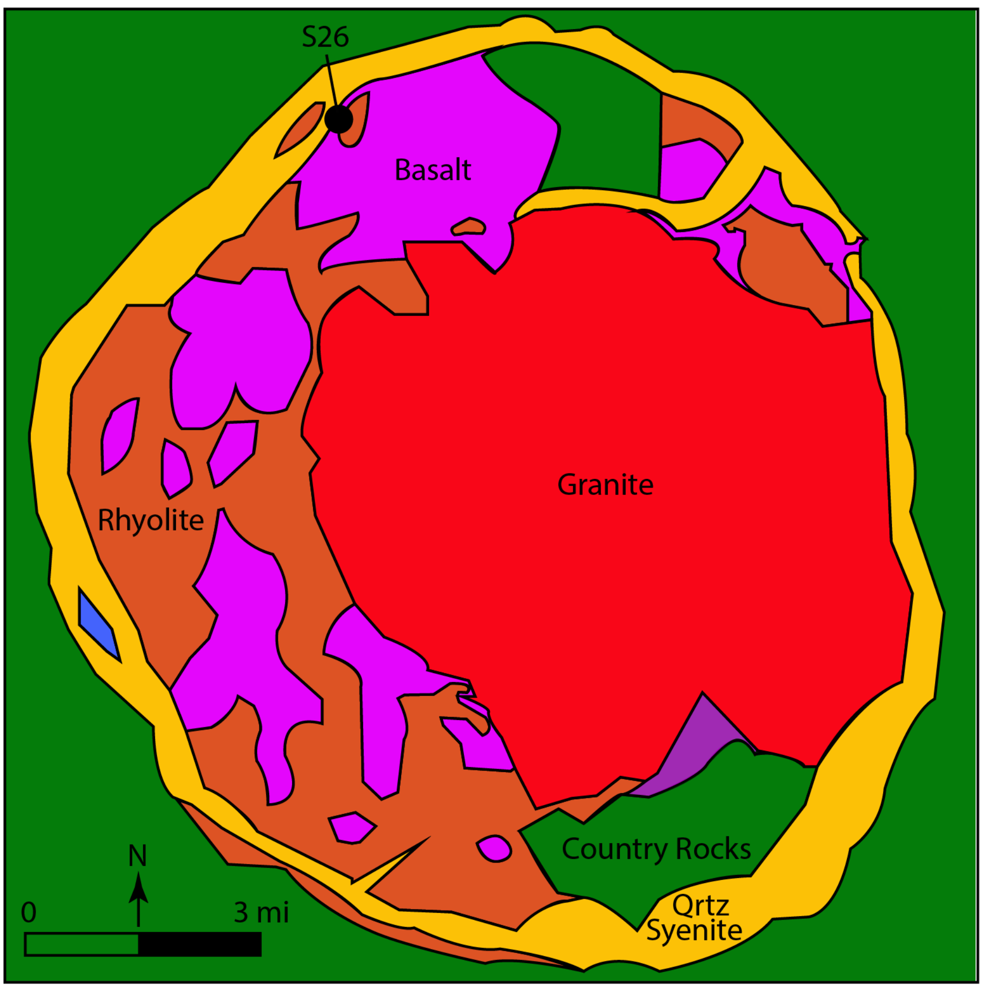


Figure 26-1. Generalized geologic map of the Ossipee Ring Complex showing the location of the Coldbrook section along the quartz syenite ring dike (After Kennedy and Stix, 2007).

The complex contains several rock types with volcanic rocks and granite in the center that are ringed by intrusive quartz syenites (rock classification given in Figure I-5). The ring dikes of the Ossipee Ring Complex formed as a magma chamber ascended towards the surface (Figure 26-2). This magma chamber was zoned, i.e, it consisted of multiple magma compositions with an upper part that was a crystal-poor rhyolitic magma, the volcanic equivalent of granite, that overlaid a more crystal-rich magma of similar composition. The base of the system contained basaltic magma. When the overlying crust was thin enough, it broke and subsided down into the chamber, producing major circular faults that mark the edges of the chamber. Magma from the chamber ascended up the faults, forming the prominent ring dike found here in the Ossipee Mountains. What is particularly interesting at the Ossipee Complex is that the magma ascended completely though the ring faults to erupt on the surface. As the overlying crust subsided downwardly into the magma chamber, a surface depression formed which filled with erupting lavas. In New Hampshire, these volcanic rocks are referred to as the Moat Volcanics that filled the deepening caldera formed by the collapse. The outcrops at Coldbrook provide glimpses of the subvolcanic processes of the complex as the various magma types mixed during turbulent ascent up the ring fractures. As the overlying crust subsided into the chamber, it disrupted the stratified magma chamber and forced multiple magma types up into the ring dike. The section at Coldbrook shows mixing between rhyolite and basalt as well as the main rock type of the ring dike, quartz syenite.

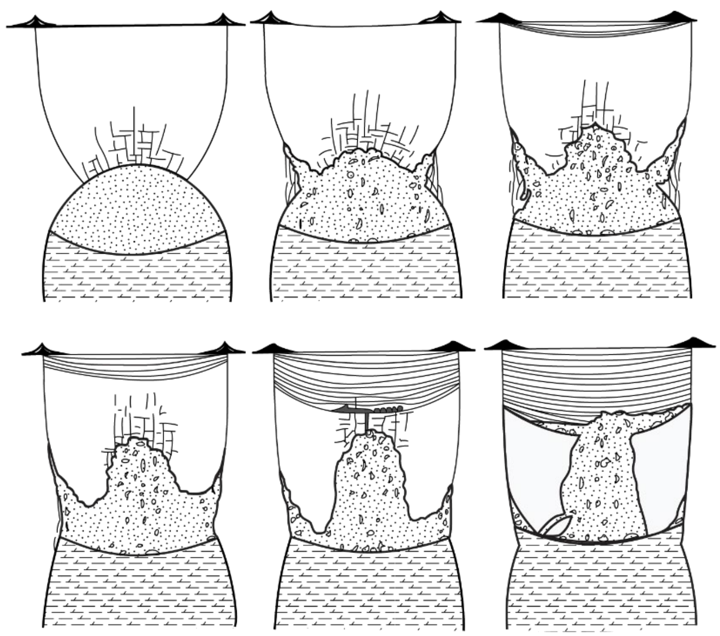


Figure 26-2. Model of the emplacement of the Ossipee Ring Dike Complex (After Chapman, 1976).

Driving Directions

From the town of Center Harbor, follow RT 25 east for 12.6 miles just past the post office in South Tamworth (Figure 26-3). Drive 0.6 miles up Mountain Road on the south side of RT 25 to a bridge crossing Coldbrook stream. Park on either end of the bridge.

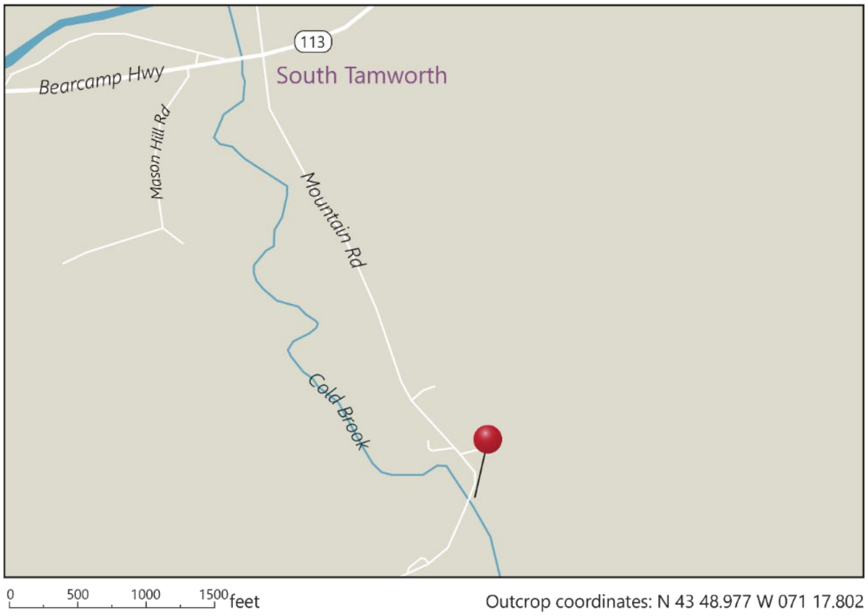


Figure 26-3. Map of the South Tamworth, NH area showing the outcrop location at Coldbrook Stream.

Walking Directions

Descend to the south side of the bridge to the outcrops exposed in the river (Figure 26-4)

Figure 26-4. View from the bridge over Coldbrook looking south. Various rock types of the ring dike are exposed here. N43°48.977', W071°17.802'

The rocks of this outcrop are mainly fine-grained volcanic rocks that lack abundant phenocrysts. Some of the basalts exposed here are exceptions, containing fairly abundant plagioclase phenocrysts (Figure 26-5). Many of these plagioclase crystals have been hydrothermally altered to greenish epidote. In spite of the overall fine grain size, the rocks are interesting because they show several features indicative of their derivation by volcanic eruptions. Figure 26-6 shows mingled rhyolite and basalt. The rock displays a crude layering formed as the two magmas of differing viscosities mixed. Stirring of the two different magma types occurred as they were forced upward in the ring dike, producing the mixed appearance seen here. The relatively high viscosities of the two magma types prevented them from completely mixing to one hybrid magma.

Other crops are fragmental basalts (Figure 26-7). These form as the outer rinds of the lava cool but because the mass of the flow is still mobile, the rind breaks apart and tends to fall of the front of the flow, only to be incorporated into the flow as it overrides its debris.

Figure 26-5. Photo of typical basalt below the bridge at Coldbrook Stream.

Figure 26-6. Mingled rhyolite and basalt.

Figure 26-7. Fragmented basalt.

Other Outcrops

Short hikes will take the interested reader to well exposed volcanic rocks in other locations in the Ossipee Mountains. The north summit of Nickerson Mountain consists of porphyritic rhyolites of the Moat Volcanics. Some locations show flow banding in the rhyolites. Basalt is also common on the mountain. Columnar basalts are nicely preserved on the way up Bald Mountain on the Bald Mountain Connector Trail off the Shannon Brook Carriage Road. The most extensive outcrops of Moat Volcanics are on the Moat Mountains near Conway that erupted from a similar volcanic system.

SUMMARY OF MAINE GEOLOGY

The geology of Maine shares much in common with New Hampshire, but there are some significant differences. The same belts of rocks in New Hampshire extend into Maine, but the metamorphic grade is considerably lower in Maine. Equivalent rocks in New Hampshire are at amphibolite to granulite facies (Figure I-6) that decrease to amphibolite to greenschist facies from western to central Maine to even lower grades in far northeastern Maine. Erosion has revealed a diagonal slice through the crust, from shallow crustal levels in northeastern Maine to greater depths all the way to Connecticut. One great advantage of the lower grade rocks in Maine is that we can examine the rocks before intense deformation and melting destroyed the original sedimentary structures. These structures allow determination of the direction of transport of the sediments, revealing that the older metasedimentary formations in western Maine were derived from the west as a result of erosion of the Taconic Mountains (e.g., the Rangeley Formation of Stop 27). One of the youngest metasedimentary formations, the Carrabassett Formation of Stop 29, is the first formation that was derived from an easterly source, indicating the arrival of a microcontinent Avalonia and the impending Acadian Orogeny.

Another aspect of Maine geology not seen in New Hamsphire is that additional belts of rocks that were accreted outboard of those in New Hampshire. These belts contain rocks that formed on the opposite side of the Iapetus Ocean that were accreated to the Laurentian margin during the Acadian Orogeny about 400 million years ago. This includes the Falmouth-Brunswick/Casco Bay island arc complex (Figure MS-1), an island arc much like the the Taconic island arc that produced the Ammonoosuc Volcanics of Stop 12, but this arc formed over a subducting oceaninc plate hundreds if not thousands of miles away from the Taconic arc. Outboard of the Falmouth-Brunswick/Casco Bay arc is another trough of sediments, the Merribuckfred Basin, named after the Merrimack, Bucksport, and Fredericton belts. Figure MS-1 illustrates that the sediments deposited in the Merribuckfred Basin were sourced from both the Falmouth-Burnswick/Casco Bay island arc, having been derived from the west, and from the Ganderian microcontinent, having been shed from the east.

Intrusive into this outermost belt of rocks in Maine are plutons of the Coastal Maine Magmatic Province (Figure MS-2). The plutons were intruded grouped during two time periods, the older between 420 Ma to 400 Ma, and a younger 380-360 Ma suite. These plutons exhibit some of the best evidence of mixing of magmas in New England (e.g., Stop 37).

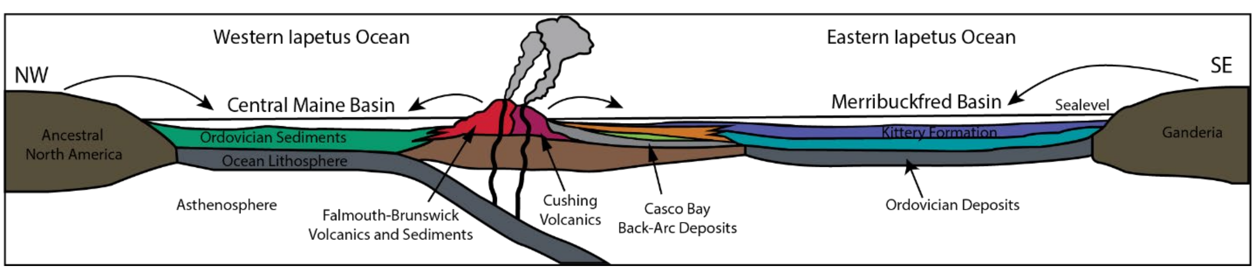


Figure MS-1. Cross section of southwestern Maine during the Ordovician, showing the Laurentian margin to the left, and two sedimentary basins (Central Maine and Merribuckfred troughs) separated by the Falmouth-Brunswick/Casco Bay island arc. Arrows show direction of sediment transport (After Hussey, 2015).

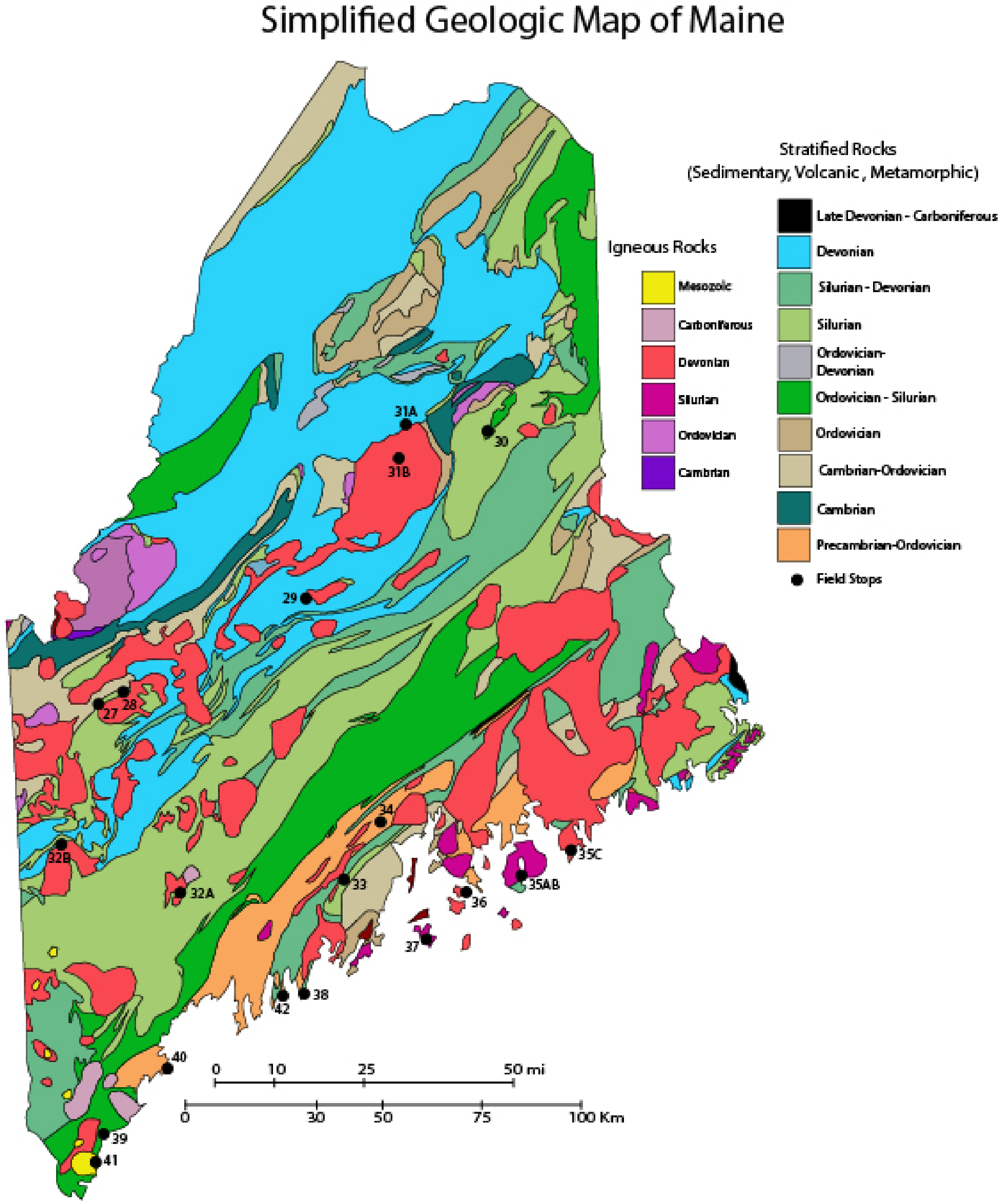


Figure MS-2. Simplified geologic map of Maine (after Marc Loiselle, 2002, simplified bedrock geologic map of Maine: Maine Geological Survey, Open-File Map SBGMM, 11" x 17" color map, scale 1:2,000,000. *Maine Geological Survey Maps*. 24. http://digitalmaine.com/mgs_maps/24)

27. Rangeley Formation, Central Maine Trough, Rangeley, ME

Following the Taconic Orogeny, vast quantities of sediment were shed from the highlands to the west of New Hampshire and Maine into a trough that was adjacent to the Laurentian margin. This trough, named the Central Maine trough after its current location, was a basin located between the Laurentian margin and the approaching microcontinent of Avalonia (Figure O-9). Sediments shed from the west off the Laurentian margin constitute the lowest and oldest formation, the Rangeley Formation, of the Central Maine trough. Above the Rangeley Formation are the Perry Mountain, Madrid, and Smalls Falls formations. With the arrival of Avalonia, sediments from the east were deposited on the older, westerly derived sediments (see the Carrabassett Formation of Stop 29). The collision of Avalonia with the Laurentian margin caused the Acadian Orogeny and the interlaying sediments of the Central Maine trough were caught in a geologic vise, subjecting the sediments to higher temperature and pressure conditions, metamorphosing, folding, and faulting the rocks. Farther to the southwest in NH and southern New England, the high grade metamorphism partially melted the sediments (See the Rangeley Formation of Stop 15). Here in western Maine, the metamorphic grade was not nearly as high as that experienced by the same formations to the southwest, allowing us to examine the original sedimentary features. These rocks are particularly interesting because the sediments are very coarse-grained, forming a rock called conglomerate. A conglomerate is a sedimentary rock with clasts or fragments that are larger than 2 mm in diameter. This outcrop has clasts that typically far exceed 2 mm.

Driving Directions

From the U.S. Post Office in Rangeley, ME, drive SE on RT 4 for 2.6 miles (Figure 27-1). Immediately beyond Overlook Road, park and walk across the road to the road cuts on the left (Figure 27-2).

Figure 27-1. Map showing the location of this stop along Route 4, east of Rangeley, ME.

Figure 27-2 showing the road cuts along the north side of the road.

On the Outcrop

These road cuts provide a wonderful example of the lower portions of the Rangeley Formation. This member of the Rangeley Formation is conglomeratic, containing up to 10 inch long clasts of a variety of sedimentary, plutonic and volcanic rocks that were shed from highlands to the northwest. Conglomerates that consist of clasts of a single rock type are called oligomict conglomerates; those with clasts of multiple rock types are polymict conglomerates as seen here. Most of the clast are well rounded, having been abraded during transportation (Figures 27-3 and 27-4). Other clasts are slabby and elongated (Figure 27-5).

It is informative to compare the clasts in the conglomerate in these photographs. Note the variability in particle size. Figure 27-5 shows large clasts up to 6 inches in length. Portions of the conglomerate are so rich in clasts that the clasts touch and form a framework; these are clast supported conglomerates. The relatively large clast size and the angularity of some of the clasts indicates that the transportation distance of the sediments was not very far. In contrast, the clasts to the right side of Figure 27-5 are much smaller and are separated by finer-grained matrix sands. These are matrix supported conglomerates. Some geologists use the term diamictites for matrix supported conglomerates. The stark difference in grain size of the clasts in the two photos indicates variable water velocity as the sediments were being deposited.

Comparing the rock types that form the clasts with outcrops to the northwest indicates that many of these clasts were eroded from outcrops only several tens of miles away. Figure 27-4 shows a granitic clast that matches the rocks of the Attean Pluton located about 30 miles to the northeast of Rangeley. Conglomerates with plutonic igneous clasts are rare because the granites tend to disaggregate into sand-size particles during transportation. Their preservation in the Rangeley Formation conglomerate attests to the proximity of the source region of the clasts.

Figure 27-3 (left). Note the coarse-grained clasts in the Rangeley Formation conglomerate (N44º55.923' W 070º 37.401').

Figure 27-4 (right). Granitic clast in the Rangeley Formation conglomerate.

Figure 27-5. This photograph shows the considerable range in clast sizes in the conglomerate, from greater than 6 inches to small, sand-sized particles (N44º55.923' W 070º 37.392').

28. Smalls Falls Formation, Central Maine Trough, Rangeley, ME

Subsequent to the Taconic Orogeny, the eastern margin of Laurentia was a continental margin with a deep Central Maine trough located between Laurentia and the approaching Avalonian microcontinent (Fiugre O-9). Sediments shed from Laurentia and Avalonia were deposited in the trough, forming the Rangeley, Perry Mountain, Madrid, Smalls Falls, and Carrabassett formations. The orientation of sedimentary structures in the oldest of these formations, the Rangeley Formation (Stop 27), indicate that the sediments were derived from sources to the west; similar structures in the youngest Carrabassett Formation (Stop 29) reveal easterly sources. Sediments that were lithified to the formations between the Rangeley and Carrabassett formations appear to have been transported along the length of the trough.

Following the deposition of the sediments of the Perry Mountain Formation, the portions of the Central Maine trough became isolated from large-scale oceanic water circulation. Oxygen became depleted in the deeper waters, probably as a result of stagnate conditions, density stratification of the water column, large input of organic matter, or from strong thermoclines, i.e, a transition layer between warmer mixed surficial water and cold, deep water. In such oxygen-poor environments, anaerobic bacteria may have been the sole living organisms; these obtained oxygen from the breakdown of sulfate (SO_4^{-2}) to release H_2S. These anoxic conditions imparted a distinct chemistry to the sediments, rendering the Smalls Falls Formation a unique appearance among the formations of the Central Maine trough.

The Smalls Falls Formation is easily recognized by its rusty appearance. The color is the result of weathering of sulfides in the rock. The high sulfur contents of the original sediments resulted from H_2S production by anaerobic bacteria as mentioned above. High inputs of organic matter also promoted anoxic conditions; hence the sediments were both carbon and sulfur rich. The carbon now is in the form of graphite, giving the rock a black color on fresh surfaces. The sulfur combined with iron to form pyrrhotite ($Fe_{1-x}S$) which, when oxidized at the Earth's surface, gives the rock its distinctive rusty color.

These rocks also contain higher concentrations of arsenic, zinc, lead, and other elements compared to other Central Maine trough formations. The combination of sulfide minerals and these elements may yield ground water with heavy metal abundances above EPA approved levels.

Driving Directions

From the U.S. Post Office in Rangeley, ME, drive SW on RT 4 for 12.0 miles to the Smalls Falls Rest Area on right (Figure 28-1). Drive to the far end of the parking lot.

Figure 28-1. Map showing the parking lot and site location for this stop at Smalls Falls Rest Area.

Walking Directions

From the parking lot, descend the stairs to the left where you'll see the scenic Smalls Falls (Figure 28-2). The falls are named after Jesse Small who lived in the area during the 1860s. Cross the wooden bridge and head up the hill to the right. At 0.1 miles up the hill are large, rusty outcrops of Smalls Falls (Figure 28-3, 28-4; N44º51.522' W 070º 31.006').

Figure 28-2. Waterfalls at Smalls Falls Rest Area with the characteristic yellow and rusty weathering of the Smalls Falls Formation.

On the Outcrop

These two photos (Figure 28-3, 28-4) illustrate the rusty weathered surfaces of the Smalls Falls Formation. The distinctive rusty color makes the Smalls Falls Formation easily recognized across a wide range of metamorphic grades. It is present farther southwest in New Hampshire and Massachusetts, where, in spite of the higher temperatures and pressures of metamorphism than here at Smalls Falls Maine, it retains its distinctive rusty appearance.

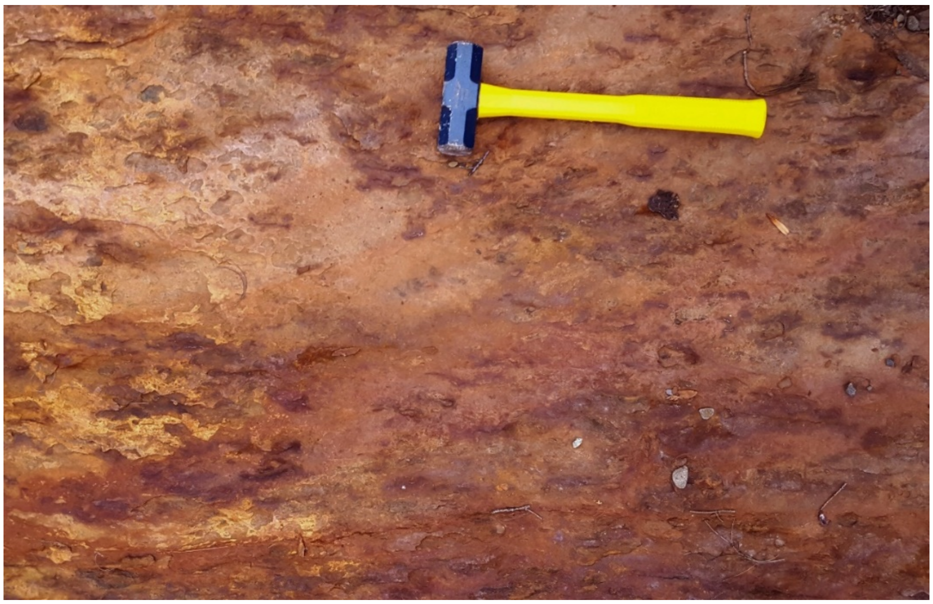


Figure 28-3. Rusty appearance of the sulfur- and carbon-rich Smalls Falls Formation.

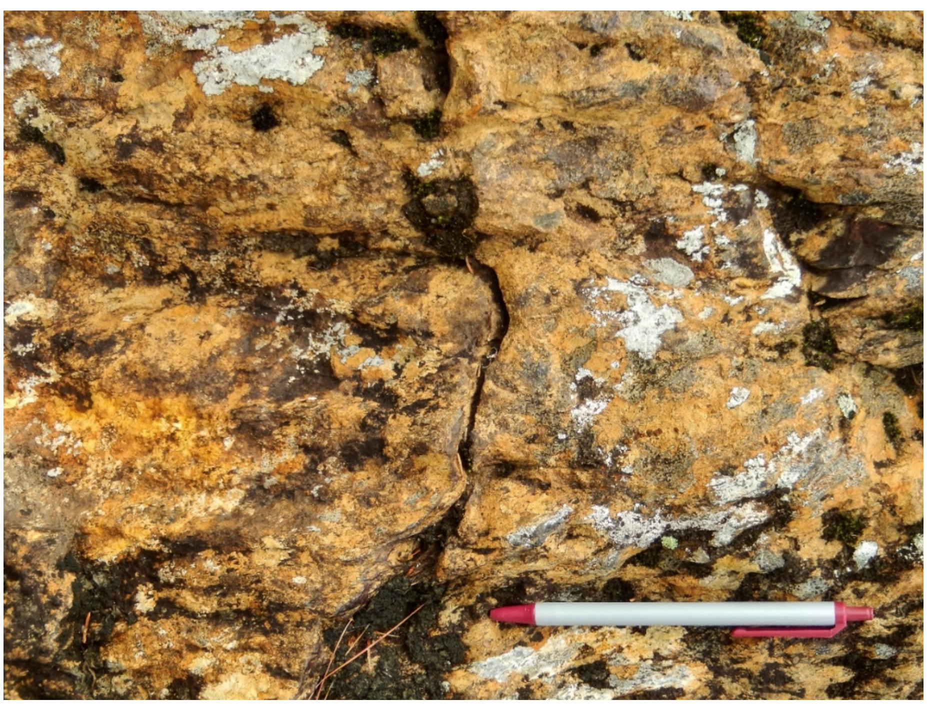


Figure 28-4. Rusty appearance of the sulfur- and carbon-rich Smalls Falls Formation.

29. Carrabassett Formation, Central Maine Trough, Big Wilson Falls, ME

Just prior to the collision of Avalon with Laurentia during the Acadian Orogeny, the decreasing distance between the two landmasses led to compression of the Central Maine trough. The trough was the depositional site of sediments shed from the eroding Taconic Mountains from the west (Figure O-9; Stops 27 and 28). As Avalon drew closer to Laurentia, sediments from this outboard terrane also were deposited in the Central Maine trough. The Carrabassett Formation represents the first, clearly discernable sedimentary package with an eastern provenance, i.e., these sediments are first that have paleocurrent indicators that they were derived from an eastern source. Here we can also examine the features of the Carrabassett Formation that indicate deposition from turbidity currents.

Turbidity currents are fast-flowing subaqueous debris flows that carry high amounts of sediment in suspension from the edge of the continental shelf down the continental slope to the ocean floor. These sediment-rich avalanches cascade down slopes under the influence of gravity because mixed sediment-water suspensions have higher densities than sediment-free water (Figure 29-1). Mud, sand, and water form a dense slurry in the currents that are probably triggered by earthquakes. Observations of modern turbidity currents indicate that they move at speeds of over 35 miles per hour, traveling with tremendous erosive power, scouring channels and canyons into the underlying sediments. The turbidity currents spread out horizontally along the floor of the basin forming submarine fans, and as the current slows down, the denser, more coarse-grained sediments are deposited first with the finer-grained sediments and muds still in suspension (Figure 29-1). When water velocities are slow enough, even the mud will settle out of the water column. Deposits from turbidity currents are therefore layered with coarse-grained sediments at the bottom and clay-rich sediments at the top of each layer. This coarse- to fine-grained characteristic enables us to determine the upward direction of the Carrabassett Formation and other turbidity deposits, i.e., which layers are oldest and which are youngest. One might think that because the law of original horizontality requires the oldest layers to be at the bottom of a sedimentary sequence, hence the younging direction should be easily determined. However, in orogenic belts, rocks are tilted and folded and the younger rocks may not necessarily be higher in an outcrop if the sequence has been tectonically flipped which is often the case in the Central Maine trough during the Acadian. Hence, turbidities provide a welcomed indicator of the younging direction. These outcrops of the Carribassett Formation are excellent examples.

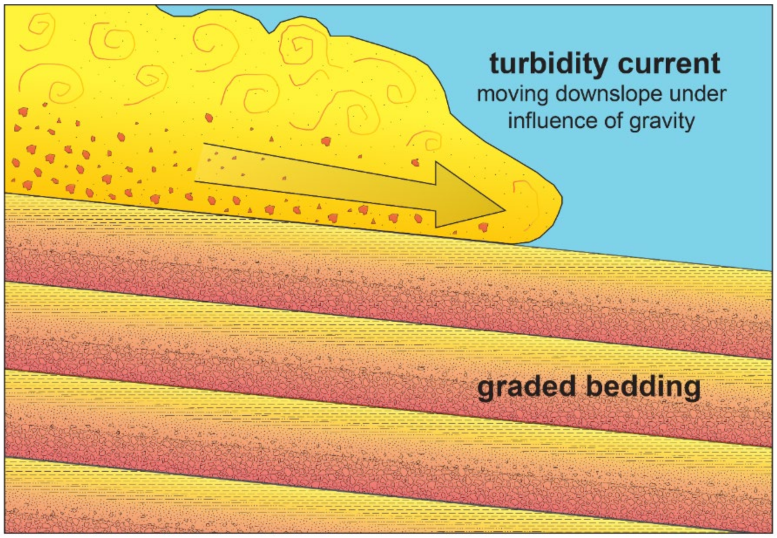


Figure 29-1. Schematic illustration of a turbidity current, showing an avalanche of sediment-rich currents sweeping downslope. (After http:/microcosta.edu/oceans/index/html)

Driving Directions

From the RT 16 and RT 6/15 junction in Abbot, ME, follow RT 6/15 north to Monson. From the municipal building in Monson, continue north for 0.5 miles and turn right on Elliotsville Road. Follow this road for 7.6 miles to the bridge that crosses Wilson Stream (Figure 29-2). Good parking spaces are available on both sides of the bridge.

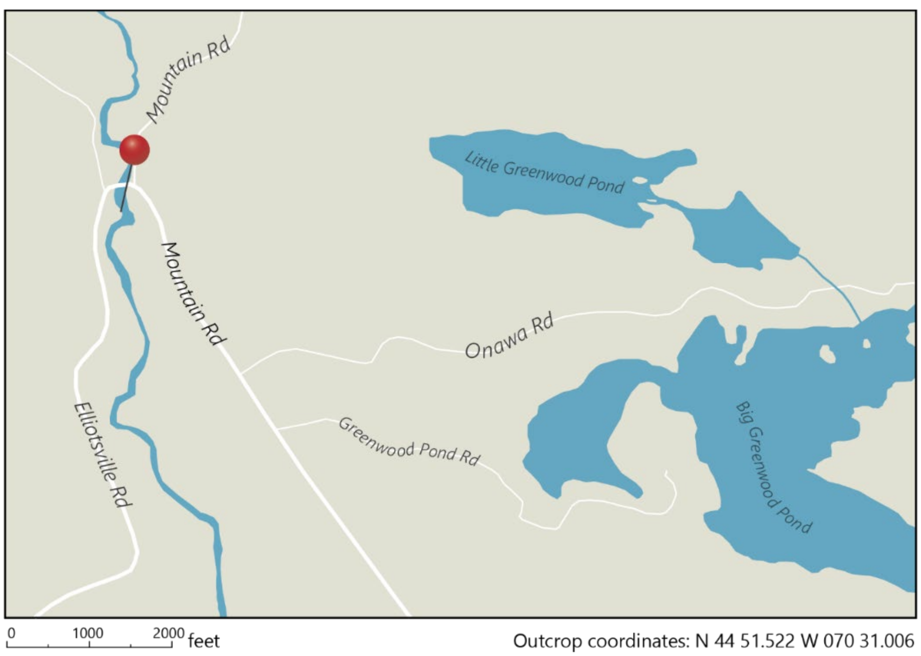


Figure 29-2. Map of outcrop location along Big Wilson Falls.

Walking Directions

Walk to the west side of the river and descend to the outcrops along the river bank Figure 29-3).

Figure 29-3. View of Carrabassett Formation outcrops along Wilson Stream (N45°22.132' W 069° 26.216').

On the Outcrop

The law of original horizontality teaches that sediments are originally deposited as horizontal sheets. The sedimentary layers at this location are no longer horizontal, having been tilted during the Acadian Orogeny. The layers seen in Figure 29-3 have been tilted downward to the right and as seen in Figure 29-6, they were also tilted almost 90° toward the viewer in this photo.

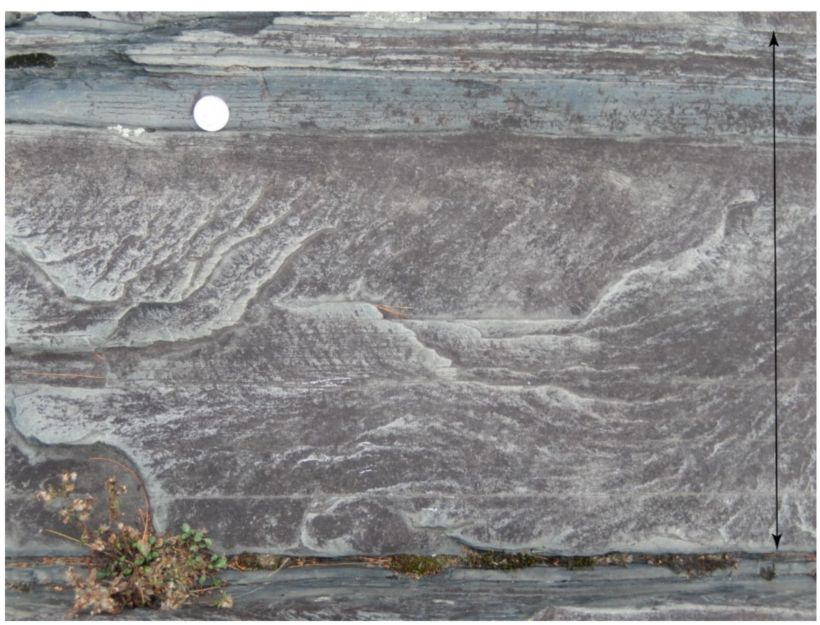


Figure 29-4. Thick bed of a turbidity deposit in the Carrabassett Formation. Arrow shows the thickness of this single bed.

Figure 29-4 shows a thick coarse-grained layer overlain by the darker, finer-grained clay-rich layer at the top of the unit. From the base of the thick layer to the top of the clay-rich layer (just above the quarter) is a single depositional unit. Commonly, the coarse-grained layer will have scoured the top of the underlying dark layer from the previous deposit (though this is not obvious in this photo but is seen in Figure 29-6). The fine-grained layer under the quarter is the youngest portion of this turbidity deposit and indicates that the sediments young upward.

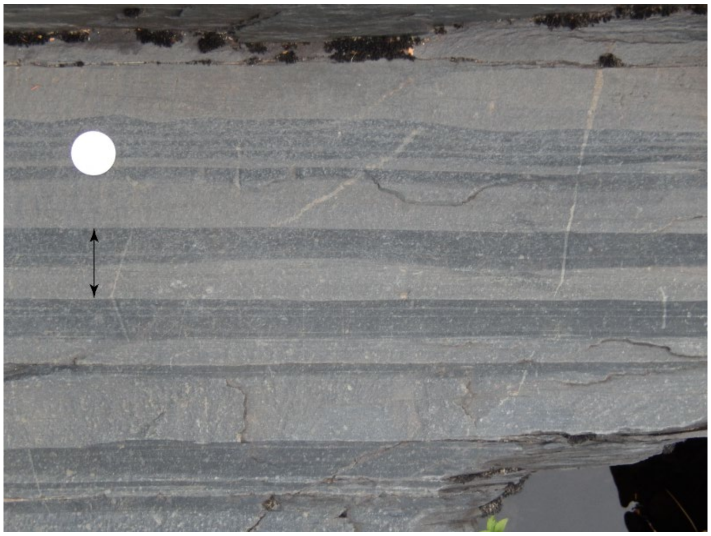


Figure 29-5. Multiple beds from turbidity currents preserved in the Carrabassett Formation. Arrow shows the thickness of a single bed.

Figure 29-5 shows successive layers of turbidity deposits. The lighter colored is coarser-grained, whereas the dark layers are finer-grained, clay-rich layers. Each light-dark layer pair represents a single depositional event with the darker gray tops overlying the lighter gray bases, schematically represented by the graded bedding of Figure 29-1.

Figure 29-6. Flute cast at the base of a turbidity current deposit in the Carrabassett Formation (N45º22.108' W 069º 26.202').

Figure 29-6 shows a horizontally elongated bulge in the rock face. This is a large flute cast or sole mark. Turbidity currents commonly carve a depression into the underlying sediments over which it flows. These depressions are elongated in the direction of current flow. Subsequent deposition fills the depression, leaving an elongated bulge as seen here. This flute cast, having a steeper slope at the up current direction (the opposite end of the cast), indicates that the current flowed towards the viewer and it marks the bottom of the bed that forms the rock wall. These features are among those used to determine that the Carrabassett Formation sediments were deposited from east to west. The other common name for flute casts, sole marks, indicting the bottom of the sedimentary layer. Hence, the sediments are younger to the left.

Figure 29-7. View of Lake Onawa from Borestone Mountain (N45° 22.633' W 069° 24.220'). The Carrabassett Formation was metamorphosed by the Onawa Pluton, converting the metasediments to erosion-resistant hornfels that form a higher relief ring of mountains around the pluton. Dashed line approximates the contact between the pluton and the Carrabassett Formation.

Another interesting observation about the Carrabassett Formation is that it forms topographic highs in this region of Maine. Wherever the Carrabassett Formation was metamorphosed by contact with an igneous intrusion, the metasedimentary rocks were baked into more weathering-resistant rocks called hornfels. These rocks form a contact aureole or zone of higher grade metamorphism around the pluton. An example of the Carrabassett Formation as hornfels is at Borestone Mountain, very near this stop at Wilson River. It's well worth the hike up Borestone where you'll see a circular range of mountains surrounding a lowland that hosts Lake Onawa (Figure 29-7). The bedrock of the lowlands is the Onawa Pluton which is much less resistant to weathering than the metamorphosed, hardened Carrabassett Formation. Farther away from the contact metamorphic effects, the Carrabassett Formation was subjected to less heat from the pluton and lacks the hardening of the contact hornfels and was also more readily eroded. The net effect of contact metamorphism of the Carrabassett Formation around the Onawa Pluton is a ring of mountains around the eroded pluton.

30. Shin Brook Formation, Shin Pond, ME

What are these European fossils doing here? Robert Neuman, the first geologist to examine the fossils at Shin Pond and related locations, must have had such a thought. The fossils at Shin Pond are unlike any North American fauna of this age. They are the same species as Early to Middle Ordovician fossils found in Ireland. So one can imagine the surprise of the first geologists who recognized that these fossils represent life forms that dwelt on the opposite side of the Iapetus Ocean. This observation begs the question: How did these fossils of Celtic origin get to North America? Islands containing these Celtic fossils were placed on the tectonic conveyor belt as subduction of oceanic crust shrank the Iapetus Ocean to the point where these islands collided with the Laurentian margin and became incorporated onto the growing continent. Faunal assemblages to the northwest of this location have Laurentian affinities. Between the Laurentian fossil localities and Shin Pond, there once existed an entire ocean. This and other outcrops in Maine and Maritime Canada were very influential in the development of the theory of Plate Tectonics in the 1960s because they proved that ocean basins are not permanent, but can be consumed by subduction.

Because of the historical significance of this fossil location, do not hammer on the outcrops: please preserve these fossils for future studies. Broken rocks at the base of the ledge are available for sampling. After visiting the fossil location, the reader might want to continue to another 1500 feet elevation to the summit of Sugarloaf Mountain for views of Mount Katahdin.

Driving Directions

Take I-95 north to RT 11 to Patten, ME. Just north of Patten, take RT 159 to Shin Pond. At the north end of Shin Pond at 9.8 miles, take a left on Wapiti Road. Drive for 1.5 miles to the gate to Camp Wapiti (Figure 30-1). Park outside the gate.

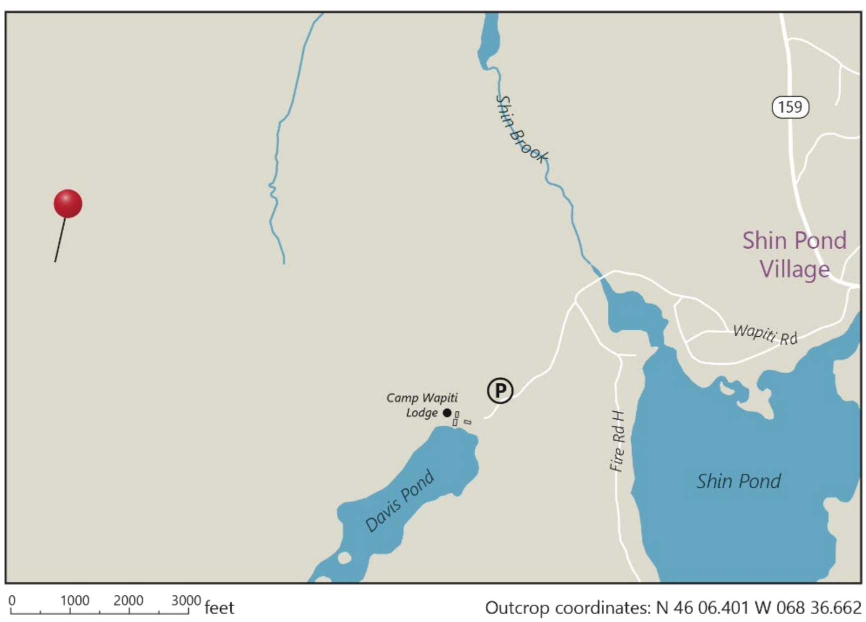


Figure 30-1. Map of parking location west of Shin Pond Village, ME.

Walking Directions

This location is so significant that I've violated the general practice of picking field locations that are easily accessible by car. But one doesn't get to see European fossils in New England every day of the week, so the site is worth the hike. Follow the dirt road that starts to the right at N46º 06.017' W 068º 34.906'. Stay on this main road, ignoring any side roads/trails. At 1.49 miles, you'll arrive at the junction with the Connecter Trail 114 at N46º05.358' W 068º 36.102'. Turn right. You should see a sign for Sugar Loaf Mountain. At 0.53 miles from this intersection, you'll arrive at another intersection at N46º05.819' W 068º 36.246'. Bear left on main road and walk for ~0.2 miles to the trail head with the Sugar Loaf Mountain Trail sign. Follow the trail for 1.33 miles where you'll arrive at a large rock (Figure 30-2) at an elevation of 1505 feet above sea level (N46º06.384' W 068º 36.681'). About 115 feet east from the rock, you'll arrive at the outcrop at N46º06.401' W 068º 36.662' at 1496 feet above sea level (Figure 30-3).

Figure 30-2 (left). Turn right into the woods at this prominent rock along the edge of the trail (N46º06.384' W 068º 36.681').

Figure 30-3 (right). The underside of this ledge contains the impressions of many brachiopods (N46º06.401' W 068º 36.662'). Brachiopods are similar to modern clams.

On the Outcrop

Brachiopods are marine animals that at first glance, look like bivalves like clams (Figure 30-4). The difference is in their symmetry. If you hold a bivalve in your hand and look at only one of the shells, flip the bivalve over and the other shell looks exactly the same. In other words, the bivalve symmetry is along the boundary between the two shells. A brachiopod had different symmetry in that the two shells are not the same, their symmetry is perpendicular to the hinge between the two shells with one shell larger than the other.

There are over 12,000 species of brachiopods identified in the fossil record and modern species are present today. Most modern species attach themselves to hard surfaces and live in cold, deep water. The majority of brachiopod species were wiped out during the mass extinction event at the end of the Permian. The brachiopods at this outcrop are Platytoechia boucoti brachiopods, a new genus and species discovered at this outcrop.

Figure 30-4. Impressions of brachiopods from the underside of the ledge of Figure 30-3.

31a. Traveler Mountain and Mount Katahdin, Baxter State Park, ME

During the Acadian Orogeny, deformation resulting from the collision of the microcontinent Avalonia with Laurentia swept across New England, starting at about 420 Ma along the Maine coast and extending to ~ 380 Ma in far northwestern Maine (see Bradley and coauthors (2000) for details). Some granitic magmas were emplaced during deformation; these, like the Bethlehem Granodiorite of Stop 16, were syntectonic, having been deformed along with the rocks they intruded. Other plutons were post-tectonic, meaning they were emplaced after the deformation front swept through Maine, and are undeformed. The granites of the Katahdin Pluton and the associated Traveler Rhyolite is a prime example.

The volcanic rocks of Traveler Mountain represent pyroclastic materials that were explosively erupted from a magma chamber. Pyroclastic flows are high-density mixtures of hot gases, magma and rock fragments whose finer-grained components are sometimes carried aloft to reach the boundary between the troposphere and stratosphere. These ash-rich clouds are dispersed by the wind, and rain ash particles over thousands of square miles. Readers old enough to remember the eruption of Mount Saint Helens of Washington on May 18, 1980 will recall that that pyroclastic eruption covered eastern Washington with several inches of ash. Not all of the magma during a pyroclastic eruption is blasted vertically; some eruptions, called nuées ardentes (French for "glowing clouds"), blast horizontally and then migrate down the volcanoes slope. These are very dangerous eruptions indeed because they move downhill at speeds of up to 100 miles an hour, burning everything in their paths. For example, in 1902, a nuée ardente from the Mont Pelle volcano on Martinique of the West Indies erupted. It enveloped the city of St. Pierre, killing approximately 30,000 people. No human inhabitants were around when the pyroclastic deposits of Traveler Mountain erupted 407 million years ago, but similar eruptions endanger human life today. As recently as June 3, 2018, a nuée ardente killed dozens of people in Guatemala.

When pyroclastic flows come to rest, they form piles of very hot material called ignimbrites or ash flow tuffs. These piles are sometimes hot enough and thick enough that the particles weld together. Such is the case with the Traveler Formation of Baxter State Park.

It is not always possible to find the eruptive vent for geologically old pyroclastic flows and to trace the transition from the erupted volcanic rocks to the plutonic rocks that mark the chamber from which the volcanic rocks erupted. The chemical composition of the Traveler Rhyolite is the same as the nearby Katahdin Granite, leading geologists to think that the two are related with the granite being the same magma as the Traveler, but having crystallized slower to form the coarser-grained granite.

Unmetamorphosed volcanic rocks are not abundant in New England, especially pyroclastic rocks. This stop enables the visitor to examine various features of pyroclastic rocks as well as features that formed because the flows were very thick, preserving compaction and cooling features. Similar rocks can be seen at Acadia National Park at Stop 35a.

This stop is another that violates the preferred "keep the stops close to the road" approach of this guide book. But the beauty of Baxter State Park is unsurpassed in New England and the hikes are a delight in and of themselves. Stop 31a takes the reader along a trail up North Traveler Mountain. Stop 31b has no specific trail to follow up Mount Katahdin, I simply recommend hiking

Katahdin and viewing the variation of grain size of the granite from the base to the summit of the mountain.

Douglas Rankin and Dabney Caldwell wrote an excellent booklet entitled "A Guide to the Geology of Baxter State Park and Katahdin" that is highly recommended for the reader who would like more detail than is presented here. The booklet comes with two maps, one of the bedrock geology, the other of surficial deposits that show the effects of Pleistocene glaciation in the park.

Driving Directions

Take I-95 north to Exit 276 at Island Falls and follow RT 159 west for 9.4 miles to Patton, ME. At junction of RT 11 and RT 159, turn right, following RT 159 for 0.3 miles. RT 159 turns left. Continue on RT 159 to entrance at Baxter State Park (At the time of writing, there is a $14 entrance fee for non-Maine residents to the park.) At 6.9 miles past the entrance, turn left into the South Branch Pond Campground road on the left. Follow for 2.1 miles to parking area on the right (Figure 31-1).

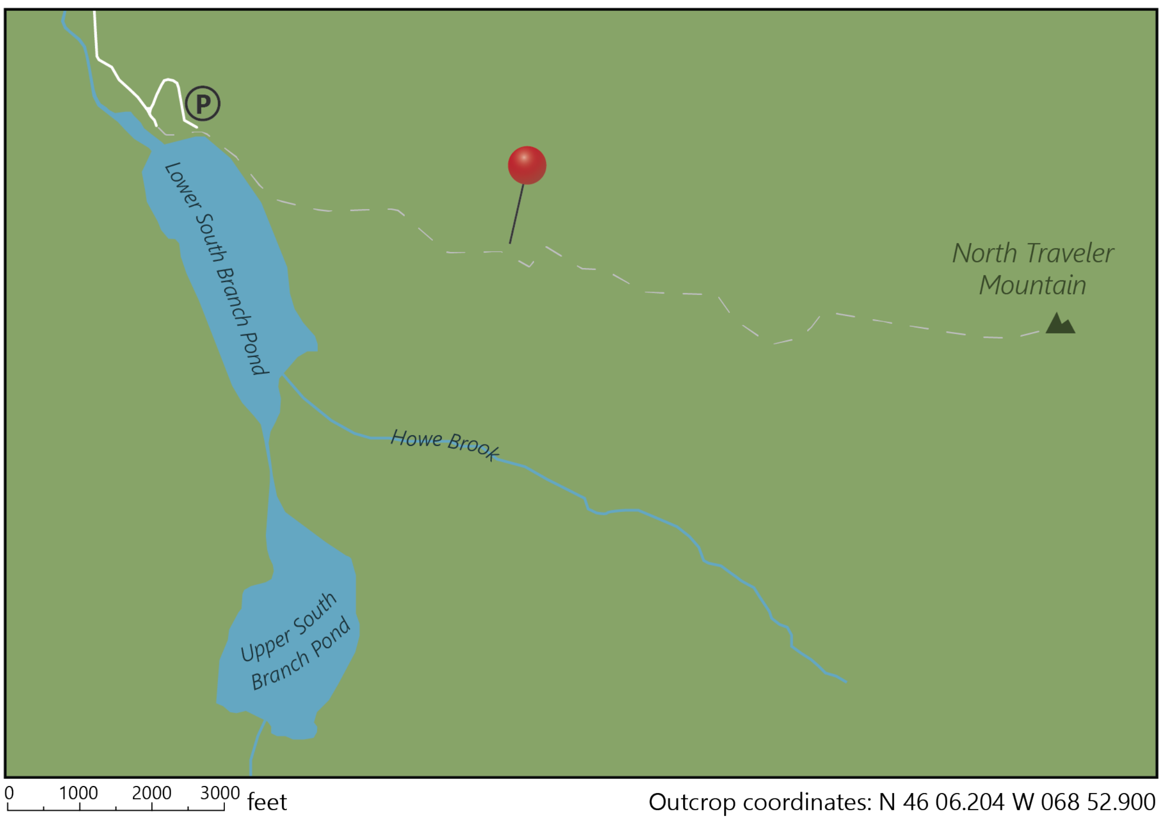


Figure 31-1. Map of parking location at the South Branch Pond Campground and site locations along the ridge of North Traveler Mountain.

Figure 31-2. View of the flank of North Traveler Mountain from Lower South Branch Pond.

Walking Directions

Begin hiking at the ranger's station at N46º 06.484' W 068º 54.015'. Proceed through the campground along the lake to the Pogy Notch Trail, 0.17 miles from the station. After 500 feet on the Pogy Notch Trail, turn left on the North Traveler Trail. At 0.7 miles there is an excellent view. Along the trail at this location, you'll see small columnar joints (at N46º 06.315' W 068º 53.213'). Also note the small vesicles, cavities formed by gas bubbles, in the rhyolite here. Larger columnar joints are at 1.07 miles at an elevation of 2310 feet (at N46º 06.204' W 068º 52.900'). Many blocks along the trail to the summit display well developed volcanic textures (Figure 31-3). Continue to 2.59 miles to the summit of North Traveler Mountain for excellent views.

On the Outcrop

Figure 31-3. Dark lenses called fiamme (Italian word for flames) are present in the Traveler Rhyolite.

Several interesting features of volcanic rocks are observable along the trail. Figure 31-3 shows dark, flattened lenses called fiamme. Some fiamme are flatted pumice fragments, others are blebs of magma of slightly different composition. Both types are flattened by compaction from the mass of the overlying tuff. When still hot, the tuff welds, or transforms from a pile of hot particles to a solid rock.

Farther along the trail you'll observe columnar jointing (Figure 31-4). The word joint is a geologic term for a fracture. Joints differ from faults in that no movement of the rocks occurs along the joint. These 4 to 8 sided columns are shrinkage fractures that formed during cooling of the magma. World-class columnar jointing is present along the southwest cliffs of Center Ridge (N46º 05.348' W 068º 53.252'). These columns range up to 6 feet thick and are several tens of feet long, dwarfing the columns seen in this photo. The adventurous visitor might want to bushwhack to visit these columns.

Figure 31-4. Columnar joints in the Traveler Rhyolite.

31b. Katahdin Pluton, Baxter State Park, ME

Figure 31-5. View from the Knife Edge to Baxter Peak, the highest point on Mount Katahdin.

Why does Mount Katahdin stand so preeminent above all other mountains in this part of Maine (Figure 31-5)? In my opinion, Katahdin is the most spectacular mountain in all of New England (this is hard to concede for a New Hampshire guy). Its rugged topography is the gift of both igneous and glacial processes.

As discussed above, the Traveler Rhyolite is thought to be the eruptive equivalent to the Katahdin Granite that crystallized in a relatively shallow magma chamber. A hike up Katahdin brings the hiker through several thousand feet of the solidified chamber, from deeper at the base of the mountain to the top of the chamber at the summit. You'll note a change in the grain size of the granite as you ascend the mountain. Near the base, the granite is fairly coarse-grained with the quartz and feldspar crystals being somewhat blocky in shape and about the same size. However, along the ridges and summits of the massif, the rock consists of fine-grained, interlocking intergrowths of quartz and Kspar called granophyre (Figure 31-6). These fine-grained rocks formed at the top of the magma chamber and the interlocking orientation of the crystals make these capping rocks more resistant to erosion than the deeper, more coarser-grained granites. Once erosion stripped away the cap rock, further erosion and glaciation more easily sculpted the granite into the great cirques seen on the east side of the mountain. The net result is the most "western-looking" mountain in New England.

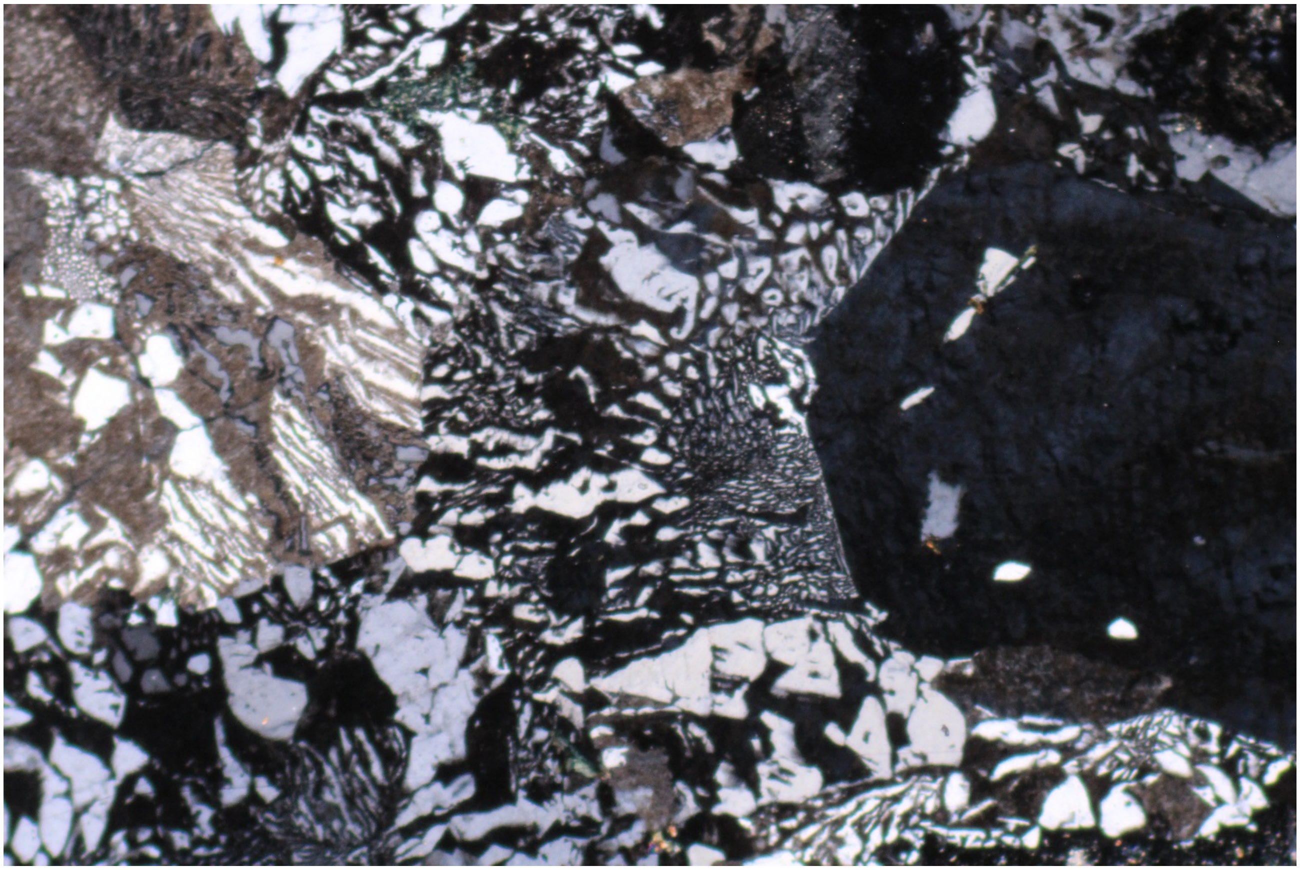


Figure 31-6. Crossed polarized light photomicrograph of the Katahdin Granite showing fine-grained, interlocking quartz and feldspar crystals called granophyric texture. These interlocking crystals make the higher elevations of Mount Katahdin more resistant to erosion than the granite at lower elevations. Field of view is 2 mm.

32a. Mount Apatite Pegmatites, Auburn, ME

Pegmatites are arguably the most beautiful of all rocks. Pegmatites are coarse grained igneous rocks, usually of granitic compositions, with crystals typically in lengths of 1 inch or more. They are characterized by the unique graphic granite texture, intergrowths of quartz and feldspar that you'll see in abundance at this stop. In fact, the name pegmatite is derived from the Greek word πήγνυμι (*pegnymi*), which means "to bind together" after this distinctive texture. This texture alone is sufficient to fascinate geologists, but it's the presence of uncommon minerals, sometimes of gem quality, that fascinate everyone. Maine pegmatites are particularly known for their gem quality tourmalines. Tourmaline and other of the more uncommon minerals in Maine pegmatites contain lithium, beryllium, cesium, boron, phosphorous, and tantalum and can serve as economic deposits for these elements. Additionally, pegmatites are excellent sources of industrial minerals. Feldspars are used for ceramics and glass, quartz for glasses and semiconductor chips and solar cells, micas, particularly muscovite, for cosmetics, insulators engineering resins etc. Pegmatites that are differentiated enough to form spodumene and petalite provide lithium for batteries and pharmaceuticals. Thus, not all pegmatites need to contain gem quality minerals to be economically valuable. These quarries were developed for the feldspars used for ceramics in the early 1900s that were processed at a mill in Littlefield Station in Auburn. The operators of the quarries were so concerned about maintaining the supply of feldspar to the mill that they disregarded pockets that contained tourmaline etc., thinking that pausing to collect the tourmalines would hinder output at the mill. One person's treasure is another person's junk.

Mt. Apatite is named after the distinctive purple colored apatite ($Ca_{10}(PO_4)_6(OH,F,Cl)_2$) found at the Pulsifer Quarry (Figure 32-1).

Figure 32-1. Photograph of a purple apatite crystal found at the Pulsifer Quarry at Mt. Apatite. (Courtesy of Myles Felch, Maine Mineral & Gem Museum.)

Driving Directions

From the junction of Routes 4/100/202 and Routes 11/121 in Auburn, head west for 1.9 miles on route 11/121 (Minot Ave). Then turn right on Garfield Road and proceed for 0.5 miles and turn left on Mt. Apatite Road (opposite Stevens Mill Road). Drive between the two baseball fields and park on the right (Figure 32-2).

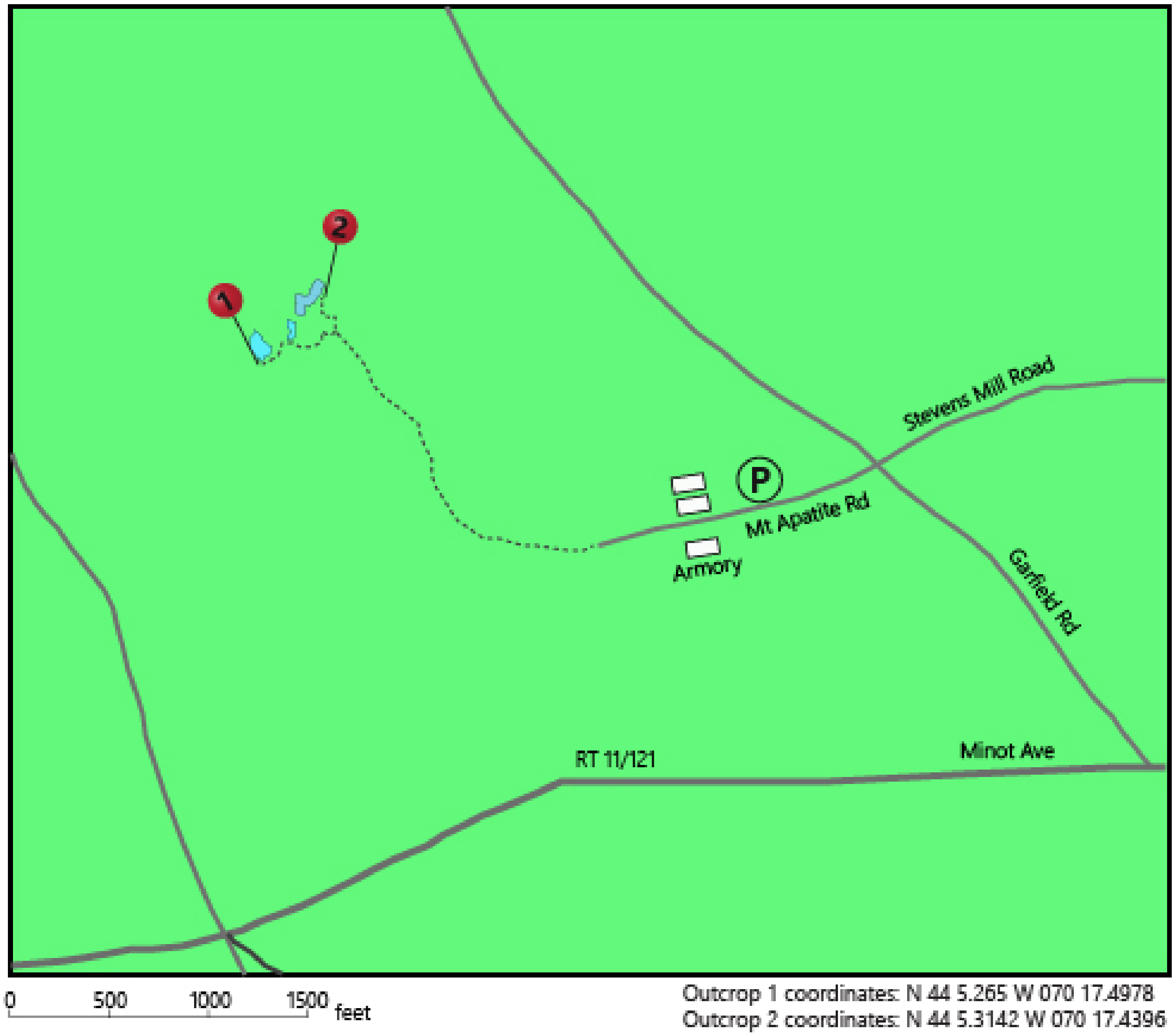


Figure 32-2. Map of the Mt. Apatite quarries. Park at N44 5.13786, W070 16.73418.

Walking Directions

From the parking lot, turn west and proceed past the gate shown in Figure 32-3. At 0.55 miles from the gate, note the trail to the left that leads to a large dump. This might be a good spot to browse after you've seen the rocks at the quarries. Searching for minerals here requires some patience because anything obvious on the surface has already been collected. Digging with hand tools is permitted, but park rules limit the depth to two feet. Continue on the dirt road to 0.67 miles where the road branches (Figure 32-4); stay left. Walk past the stone wall on the left to the Maine Feldspar Quarry at 0.71 miles. You'll see a quarry filled with a small pond. Walk left of the pond to 0.76 miles to a larger quarry and pond. At 0.82 miles, you'll find many blocks of good pegmatite showing graphic granite texture (Figure 32-5; N44 5.265, W070 17.4978). Also present is garnet-bearing pegmatite (Figure 32-6). When finished here, return to the stone wall and turn left to the Greenlaw Quarry. Here you'll find fairly large black tourmaline crystals in blocks (Figure 32-7. N44 5.3142, W070 17.4396).

Figure 32-3. General view of the start of the hike to the Mount Apatite quarries. The mileage for this visit starts at the gate shown in the right portion of the photo.

Figure 32-4. At 0.67 miles, the trail diverges. The left branch proceeds to the Maine Feldspar Quarry, the right branch to the Greenlaw Quarry.

At the Outcrop

At this location, you'll find many blocks of pegmatite with coarse-grained crystals, many of which are intergrown to form graphic granite textures (Figure 32-5A & B). Introductory geology classes teach that large crystal sizes in igneous rocks require very slow cooling rates at great depths in the Earth's crust. This generalization does not apply to pegmatites, the coarsest grained of all igneous rocks, because they are commonly present at moderately shallow depths where they were intruded into relatively cold country rocks. Hence, the process that forms large crystals and intergrowths of quartz and feldspar in pegmatites has fascinated geologists for decades. Fortunately, experimental petrology has greatly enhanced our understanding of pegmatites.

For a crystal to form and grow from a magma, it first needs to nucleate, i.e., incipient arrangement of atoms must occur to form a starting point from which the crystal can grow. The formation of nuclei and initiation of crystal growth under equilibrium conditions ideally starts when a magma cools to the liquidus or the initial crystallization temperature of a magma. But because nucleation is sluggish, crystals may not readily form at the liquidus; instead nature sometimes needs to give the nuclei a kick in the backside to promote growth. One of the best kicks is when the magma cools below the initial crystallization temperature, that is, it cools at a slightly faster rate than that which the nuclei can form at equilibrium conditions at the liquidus (but not fast like a lava flow or a very shallow intrusion). This dropping below the liquidus temperature to initiate crystallization is called undercooling. At a certain undercooling rate, not too fast or too slow, it is as though the nuclei suddenly realize that they should have been growing and now grow more quickly in order to catch up to where they should have been. In these special volatile-rich pegmatitic magmas, there are few crystal nucleation sites, so when they start to grow to form crystals, there are few neighboring crystals to compete for space and each crystal can grow unimpeded to large crystal sizes. The unique pegmatite magmas allow rapid diffusion of the essential elements for these rapidly growing crystals.

During specific undercooling conditions, both quartz and feldspar undergo rapid grown in an attempt to reach equilibrium conditions with the magma, and in the process, race each other in growing out into the magma. As they race, the crystals intergrow, forming the unique pegmatitic texture, graphic granite as seen in Figures 32-5A and B. When viewed on end (Figure 32-5B), the crystals resemble cuneiform writing, hence the name graphic granite.

Figure 32-5A. Lengthwise view of intergrown feldspar (white) and quartz (gray) showing graphic granite texture that is characteristic of granitic pegmatites. B. End view of graphic granite showing the resemblance to cuneiform writing.

Figure 32-6. Cluster of garnet in the pegmatite at the Maine Feldspar Quarry.

Figure 32-7. Black schorl tourmaline at the Greenlaw Quarry.

So how large do crystals in pegmatites get? Those seen at these quarries are typical but are nowhere near the large crystals found elsewhere in Maine and around the world. The world's largest beryl ($Be_3Al_2Si_6O_{18}$) was 49 feet long and weight 419 tons. Crystals of spodumene ($LiAl(SiO_3)_2$) up to 47 feet, quartz up to 20 feet, muscovite ($KAl_2(AlSi_3O_{10})(F,OH)_2$) 15 feet in diameter, and microcline crystals ($KAlSi_3O_8$) 75 feet long are reported in the literature. Crystals of these sizes are exceptional, but the more generic sizes found in this stop are still very impressive.

Many pegmatites are referred to as simple, being dominated by feldspars and quartz with few if any more exotic minerals that are of economic interest. Complex pegmatite are zoned from their margins to their cores (Figure 32-8) and may contain gem quality minerals in their interiors. Complex pegmatites tend to be zoned from feldspar-rich margins to more quartz-rich interiors. Some of these complex pegmatites are zoned to open pockets in their interiors, pockets that can approach small room size. It is into these open pockets that once contained hydrothermal fluids that gem quality minerals grew. When discovered, these pockets can be vacant, forming open rooms, or may be filled with clays.

The minerals that crystallize into these pockets are rich in elements that are excluded from the crystal structures of previously crystallized minerals. Enrichment in the late stage magma and fluids in elements such as lithium, beryllium, cesium, boron, phosphorous, and tantalum lead to tourmaline, beryl, spodumene, lepidolite, pollucite, etc. forming. Tourmaline is particularly interesting because it can contain a wide range of elements, forming many different species of tourmaline.

The common variety of tourmaline, schorl ($NaFe^{2+}{}_3Al_6Si_6O_{18}(BO_3)_3(OH)_3OH$) is found in many pegmatites. As the magma evolves and the lithium concentrations increases, the amount of lithium in the tourmaline correspondingly increases to the point where elbaite ($Na(Li_{1.5},Al_{1.5})Al_6Si_6O_{18}(BO_3)_3(OH)_3OH$) forms (e.g., Figure 32-9). These are the beautiful green to pink varieties that are so valuable.

Mica also shows an increase in lithium, eventually forming the purplish colored lepidolite ($K(Li,Al,Rb)_2(Al,Si)_4O_{10}(F,OH)_2$) that takes the place of muscovite ($KAl_2(AlSi_3O_{10})(F,OH)_2$). Some pegmatites in Maine also contain spodumene, ($LiAl(SiO_3)_2$).

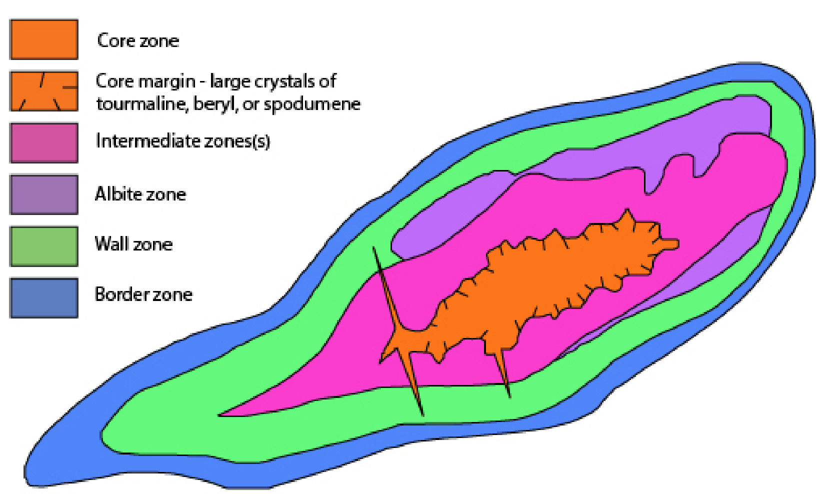


Figure 32-8. Schematic cross section of a pegmatite, showing several zones and an interior core zone where exotic minerals are found (after Bradley and McCauley, 2013).

Figure 32-9. Zoned tourmaline (schorl in the lower right to elbaite in upper left) in the Mt Mica pegmatite, Oxford pegmatite field, western Maine. Quarter for scale.

32b. Maine Mineral & Gem Museum

Since the chance of seeing a gem quality mineral at the Mt. Apatite dumps is exceedingly rare, you can satisfy that craving by visiting the Maine Mineral & Gem Museum at 99 Main Street in Bethel. The museum has an exceptional collection of gem quality minerals extracted from pegmatites across the state, as well as an excellent meteorite display. The entrance fee is modest (at the time of writing, $15 for adults, $12 for seniors and veterans, $10 for students and children, and free for children under 12), and seeing the collection is well worth the price. Make sure you don't miss the rock and mineral display in the rock garden on the grounds outside the museum. Several of the photos below are of these rocks and minerals.

Figure 32-10. Gem quality tourmaline extracted from a pegmatite in Maine on display at the Maine Mineral and Gem Museum at Bethel, Maine. Crystal length is ~ 10 inches long.

Tourmaline has a complex formula with many substitutions that give it a wide range in compositions and a plethora of names based on the compositional diversity. It's general formula is $XY_3Z_6(T_6O_{18})(BO_3)_3V_3W$, where X = Ca, Na, K, ▢ = vacancy, Y = Li, Mg, Fe^{2+}, Mn^{2+}, Zn, Al, Cr^{3+}, V^{3+}, Fe^{3+}, Ti^{4+}.

At the Greenlaw Quarry, the most common variety of tourmaline, schorl, is present (Figure 32-7). The formula for schorl is $NaFe^{2+}{}_3Al_6(BO_3)_3Si_6O_{18}(OH)_4$. As the pegmatite magma/fluid evolved in Maine pegmatites, Li contents increased and the tourmaline shows a progressive change to elbaite, the Li endmember $(Na(Li,Al)_3Al_6(BO_3)_3Si_6O_{18}(OH)_4$. This tourmaline forms the gem quality green to pink varieties that is so valued by gem collectors and jewelers. An examples is shown in Figure 32-10.

Extraordinarily large crystals of beryl have been mined from Maine pegmatites (Figure 32-11). An essential constituent of beryl is beryllium (Be), another element that is excluded in the crystal structures of feldspar, quartz, and other earlier crystallizing minerals in granites and pegmatites. Only highly evolved magmas and fluids will concentrate sufficient beryllium to form beryl. Some of the largest beryl crystal ($Be_3Al_2Si_6O_{18}$) ever found was quarried from the Bumpus Quarry in Albany, Oxford County, Maine, being up to 27 feet in length and weighing over 26 tons.

Figure 32-11. Large beryl crystal from a pegmatite in Maine. The mineral is on display at the rear entrance (east side) of the Maine Mineral and Gem Museum at Bethel, Maine.

Figure 32-12. Photo of a pegmatite with abundant green tourmaline. The rock is on display in the rock garden at the west side of the Maine Mineral and Gem Museum at Bethel, Maine.

Figure 32-13. Photograph of pegmatite on the grounds of the Maine Mineral and Gem Museum showing dark colored tourmaline in the center and purple colored lepidolite mica ($K(Li,Al_3)(AlSi_3)O_{10}(OH,F)_2$) at the left.

33. Pseudotachylite in the Cape Elizabeth Formation, Norembega Fault System, Somerville, ME

The Norumbega Fault System (Figure 33-1) is a major fault system that rivals the more famous San Andreas Fault of California in length (Figure 33-2). Is it a strike slip, right lateral fault system that began as a 30-40 km wide shear zone in the Middle Devonian, extending from Maritime Canada down through Connecticut, a distance of greater than 700 miles. A strike slip fault is one where the fault plane is vertical or near vertical and the rocks on the opposite sides of the fault move horizontally with very little vertical motion (Figure I-7C). Imagine that you are standing on one side of the fault, viewing to opposite block. During movement along the fault, the opposing block would move to the right for a right lateral strike slip fault. In this case, the coastal Maine block has moved to the southwest with respect to the block on the inland side of the Norembega Fault System moving to the northeast.

The rocks at this outcrop belong to the Cape Elizabeth Formation. In the Middle Devonian, earliest motion along the fault system deformed rocks that were relatively deep and hot. Hence, rocks in the proximity of the fault deformed ductally, i.e, the rocks were deformed by stretching without breaking and underwent grain size reduction to produce rocks called mylonites.

The youngest activity along the fault occurred in the Early Permian, giving a lifespan of ~80 million years of fault movement. Uplift and cooling of the rocks by the Early Permian changed the stress regime from ductile to brittle deformation and focused the strain to relatively narrow fault strands. Movement along these younger faults, cutting rocks that were relatively rigid by this time, produced high degrees of friction in the rocks, heating them to the point where melting occurred. Partial melts along fault surfaces are called pseudotachylite. Pseudotachylite resembles tachylite, a vitreous form of volcanic glass, and is called pseudotachylite because it is not of volcanic origin but forms by frictional melting.

What makes this outcrop (and other Norumbega Fault System outcrops) interesting is that each band of pseudotachylite represents melting produced during an individual earthquake. At other Norembega Fault System outcrops, dozens of bands of pseudotachylite have themselves been sheared during younger seismic events, indicating that the fault system, like its California cousin, was long lived. Unlike Californians, Mainers need not worry about current activity along the fault since the compressive forces that slid the coastal Maine rocks to the south west are no longer present.

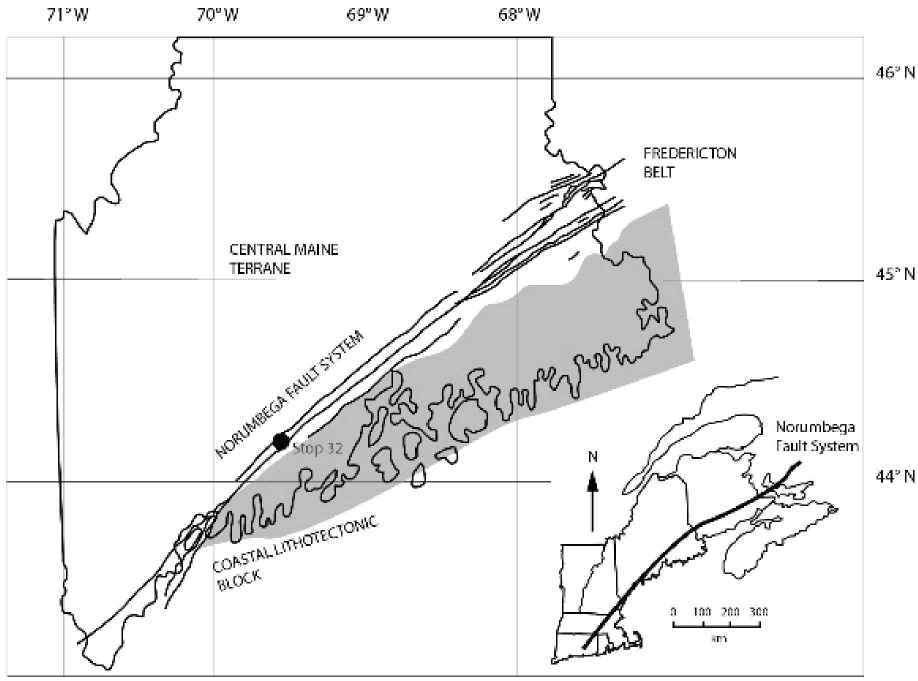


Figure 33-1. Map of New England showing the location of the Norumbega Fault System. It extends for ~ 750 miles from New Brunswick to Connecticut (After Ludman and West, 1999).

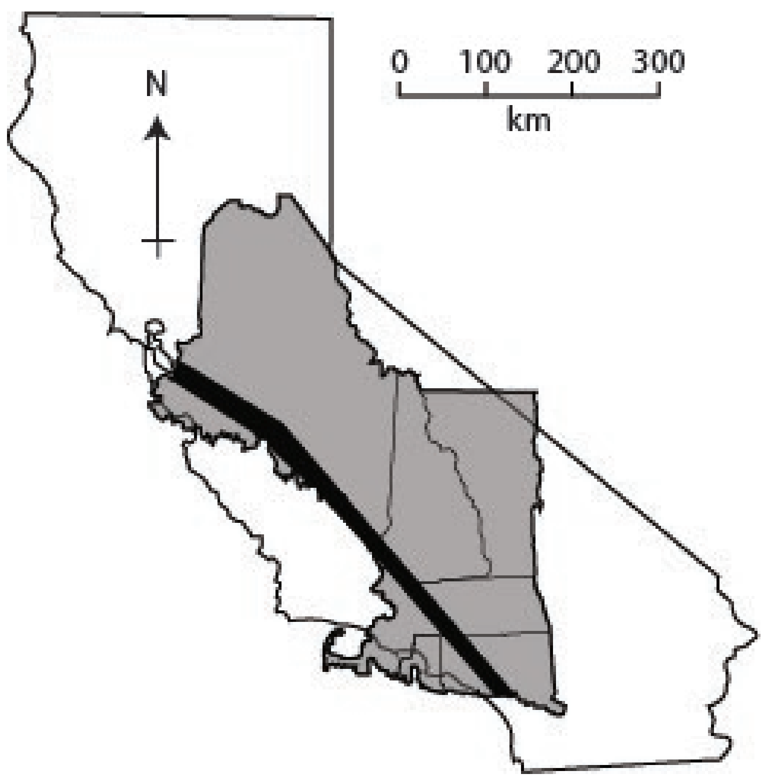


Figure 33-2. Map of New England and the Norumbega Fault System flipped and overlaid on a map of California. The Norumbega Fault System rivals the San Andreas Fault in length (After Ludman and West, 1999).

Driving Directions

From Union, ME, drive west on RT 17 past the intersection of RTs 17 and 206 (about 8.5 miles). Continue on RT 17 for 0.8 miles to the intersection of RT 17 and Jones Road; turn left on Jones Road (Figure 33-3). About 0.68 miles down Jones Road, just before a light blue trailer home on the left, there is a faint logging road on the right (N 44º 15.099’ W 069º 27.856’). Park here.

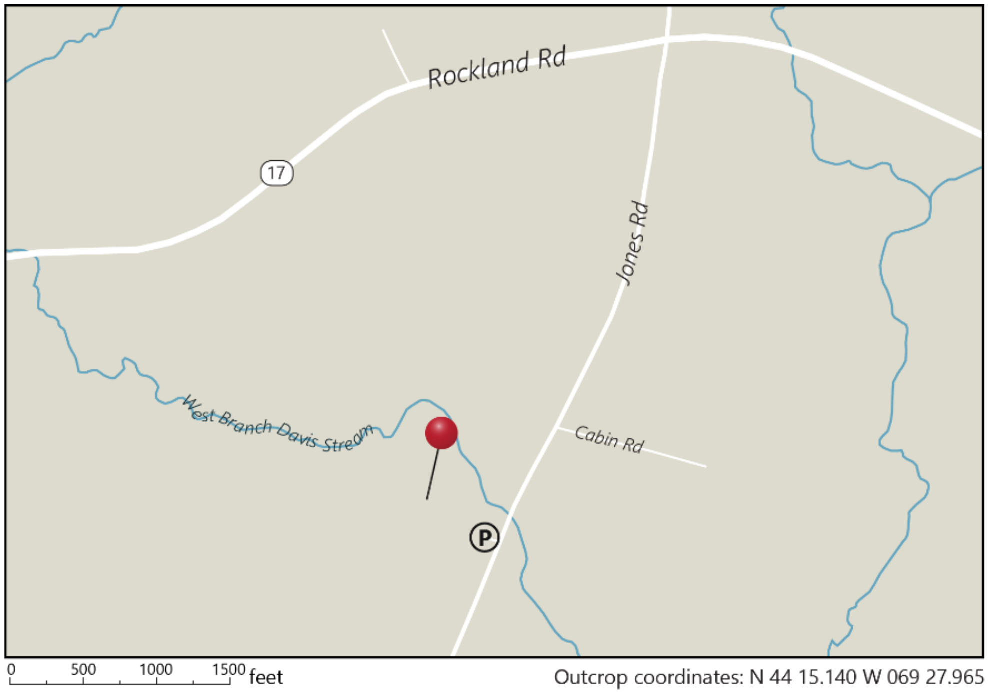


Figure 33-3. Map of the Jones Corner shear zone area near Union, Maine.

Walking Directions

The outcrop is about 175 yards up the logging road. Walk up the logging road to where it forks, take the right fork to an outcrop to your left as seen in photo Figure 33-4.

On the Outcrop

The rocks at this outcrop are mylonites, rocks that underwent grain-size reduction during movement along the fault as the stress deformed the ductile, hot rocks. As the rocks moved towards the surface over several tens of millions of years, they cooled and became more brittle. Continued motion along the fault caused frictional melting now preserved as pseudotachylite, shown in the middle of Figure 33-5. The pseudotachylite displays a finer grained margin where the melt quenched against the colder, surrounding rock, and a flow fabric in the interior from flow of the melt.

Figure 33-4. The outcrop of interest is located through the trees about 20 yards south of the logging road (N 44º 15.140’ W 069º 27.965’).

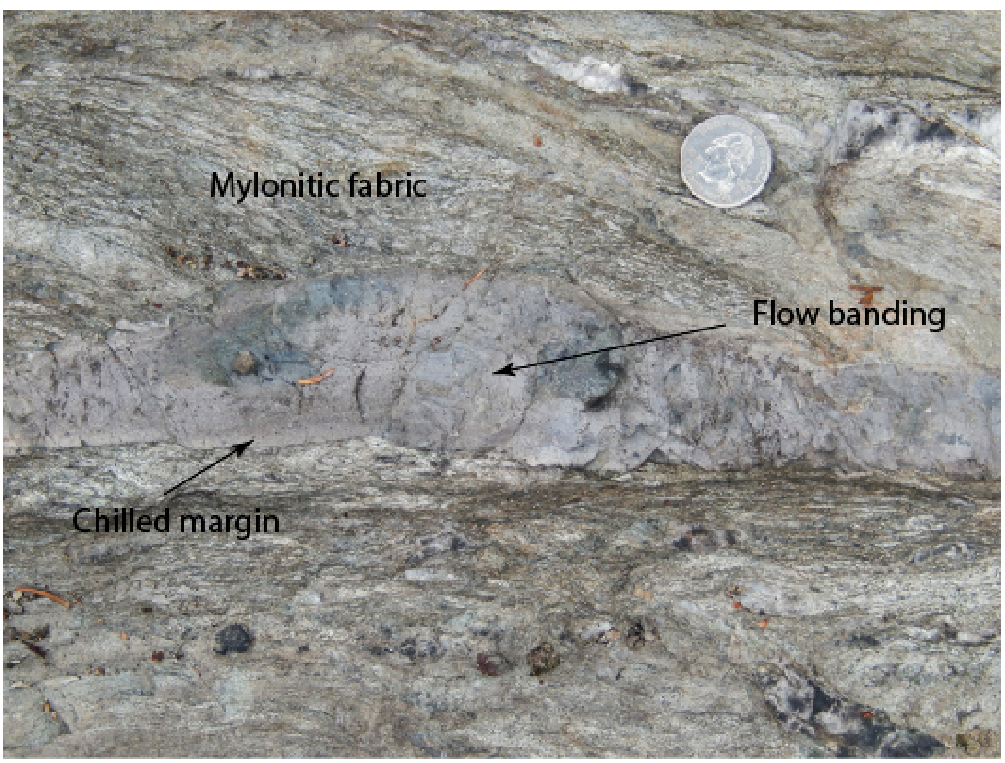


Figure 33-5. Mylonites of the Cape Elizabeth Formation and pseudotachylite bands.

34. Appleton Ridge Formation, Frederickton Trough, Liberty, ME

We've visited two troughs that served as sites of sediment deposition following the Taconic Orogeny, the Connecticut Valley – Gaspe trough with the Gile Mountain (Stop 11) and Waits River (Stop 10) formations, and the Central Maine trough with Stops 14, 15, 27, 28, and 29. Outboard of the Central Maine trough was a relatively elevated land mass call the Miramichi Highlands (similar to the Falmouth-Brunswick arc of Figure MS-1). Oceanward of the highlands was one more depositional site called the Frederickton trough (Figure MS-1). Clay-rich sediments that were to become the Appleton Ridge Formation were deposited in the trough during the Silurian Period about 430 million years ago. During Acadian metamorphism, the Appleton Ridge Formation transformed into some of the most beautiful metamorphic rocks in all of northern New England. The schist at this stop hosts abundant, up to 3 inch long, crystals of staurolite ($Fe^{2+}_2Al_9O_6(SiO_4)_4(O,OH)_2)$). While staurolite is present in other formations of northern New England, the abundance of staurolite at this location is unique. Staurolite is a metamorphic mineral that occurs as clay-rich sediments are metamorphosed to distinct pressure and temperature conditions of the amphibolite facies (Figure I-6).

These rocks are about 30 miles away from the strand of the Norubega Fault System as seen in the previous stop. Hence, the deformational effects of that fault are not as highly expressed as in the rocks at Jones Corner, but evidence of right lateral deformation is also present here, showing the widespread deformational effects of this major fault system.

Driving Directions

About 6 miles south of Liberty, ME, follow RT 220 to RT 105. Turn left (east) on Burkettville Road. At 5.1 miles, turn left on Appleton Ridge Road. At 7.2 miles, pull off on small dirt drive on right and park (Figure 34-1). Or, from junction of RT 105 with RT 131/17 in Union, ME, drive north on RT 105/131 for 3.2 miles. Turn left on RT 105. At 4.2 miles, turn right on Appleton Ridge Road. At 6.3 miles, park on small dirt drive on right.

Figure 34-1. Map showing the location of the site of interest along Appleton Ridge Road.

Figure 34-2 showing the outcrop of the Appleton Ridge Formation (N 44º 17.827’ W 069º 15.422’).

Walking Directions

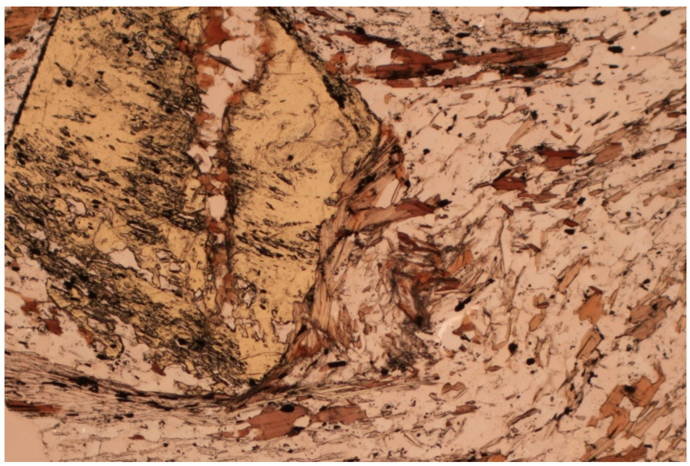
Walk east for 50 yards to outcrops in blueberry field to the outcrop shown in Figure 34-2.

Figure 34-3. Abundant staurolite crystals in the Appleton Ridge Formation.

On the Outcrop

The Appleton Ridge Formation was metamorphosed during the Acadian Orogeny and now contains a wonderful assemblage of metamorphic minerals. Most obvious in Fiogure 34-3 is the abundance of black-brown staurolite crystals. Also present but not as easily seen are garnet and andalusite. These minerals indicate that the rocks were heated to temperatures between 500 and 600° C (~932 to 1100° F). That may seem hot indeed, but not enough to partially melt the rocks like those of the Rangeley Formation at Stop 15.

Note the alignment of the staurolite crystals in outcrop. Some of the crystals are deformed, showing tapered ends. These rocks experienced post peak-metamorphic shearing, probably during the Late Devonian to Early Carboniferous time during movement of coastal Maine to the southwest along the Norumbega Fault System.



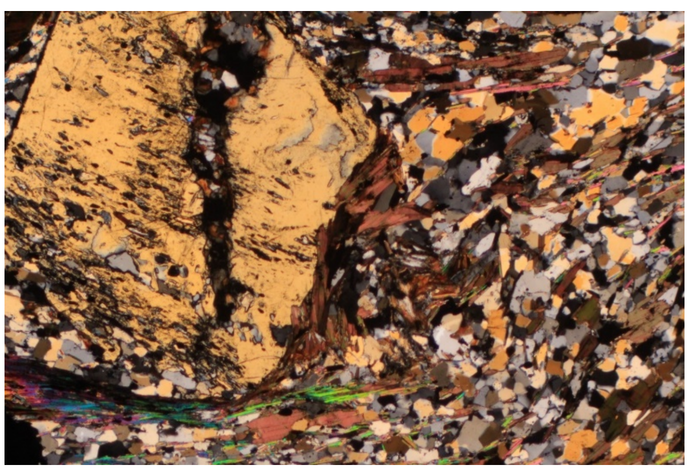


Figure 34-4. Crossed polarized light (upper photo) and plane polarized light (lower photo) photomicrographs of the Appleton Ridge Formation showing a yellow colored staurolite crystal in the left portion of the photos. The reddish brown mineral of the upper photo is biotite; the blue to green mineral of the lower photo is muscovite. The remaining white, gray, and pale yellow minerals are quartz and feldspar.

Figure 34-5. Boudinaged quartz vein in the Appleton Ridge Formation.

Figure 34-5 shows a broken quartz vein in the middle of the photo. The rocks to the right of the quartz vein moved towards the viewer whereas the left portion of the photo moved upward, away from the viewer, stretching and breaking the quartz veins. This shearing motion along the vein with the right side toward the viewer and the left side away, means the rocks experienced right lateral motion. Even though this outcrop is about 30 miles from the main portion of the Norumbega Fault System, at the time of deformation, the rocks were deep enough to experience ductile deformation.

35. Acadia National Park, ME

Samuel de Champlain sailed into Frenchman Bay in 1604 and, upon viewing Mount Desert Isle, called it “Isles des Monts Déserts” or Island of Bare Mountains. Cadillac Mountain of Mount Desert Isle is the highest point (1530 feet) along the North American Atlantic coast between Cape Breton Highlands of Nova Scotia and Mexico.

The beauty of Acadia National Park justifies its inclusion in any guide book; the geology of Acadia National Park is interesting enough to require it. The park hosts rocks that tell a story of magmas being emplaced at shallow crustal levels to volcanic rocks being erupted from those shallow chambers, each with many interesting features to see. The magmas that erupted to form the volcanic rocks also crystallized to form granites at relatively shallow depths and Acadia National Park allows the visitor to view the transitions from volcanic to shallow plutonic settings. Additionally the dikes at Schoodic Point represent the first steps of continental rifting that led to the development of the Atlantic Ocean basin. Combining the beauty of the Maine coast with a fascinating geologic story is a winning combination.

Several plutons are present in the park, the most famous is the Mount Cadillac Granite that underlies most of Mount Desert Island. This and other plutons along the Maine coast constitute the Coastal Maine Magmatic Province, a belt of two groups of plutons with ages ranging from and older group between 430 to 400 million years old and a younger group with ages from 380 to 360 million years. Both groups of plutons intruded rocks of the Ganderia terrane, a microcontinent that originated on the opposite side of the Iapetus Ocean (Stop 21). Ganderia docked with Laurentia during the Salinic Orogeny about 430 million years ago. Other plutons visited at Stops 35b, 36, and 41 also belong to the Coastal Maine Magmatic Province.

I highly recommend that visitors who desire more detail about the geology of the park purchase the booklet “The Geology of Mount Desert Island: A Visitor’s Guide to the Geology of Acadia National Park” by Richard Gilman and coauthors. The booklet is available from the Maine Geological Survey web site. The book comes with two maps, one showing the bedrock geology of the park, the other the surficial geology showing the effects of Pleistocene glaciation of the region. Another good book is “Guide to the Geology of Mount Desert Island and Acadia National Park” by Duane and Ruth Braun.

35a. Cranberry Island Series Volcanics

The rocks of the Cranberry Island Volcanics are abundant on the Cranberry Islands offshore from the park. But they are also well exposed to the west of Southwest Harbor. As with the ash flow tuffs of the Traveler Rhyolite (Stop 31a), the Cranberry Island Volcanics are also pyroclastic rocks. What makes these rocks interesting is the large amount of angular fragments present in the tuff. Pyroclastic eruptions are violent events; pumice fragments and rock fragments broken off older rocks surrounding the volcanic vent are present as well. The volcanic rocks were erupted as pyroclastic flows and lava flows similar to those at the Ossippee Mountain complex of Stop 26 (Figure 26-2).

As the magma chamber emptied during these pyroclastic eruptions, radial fractures formed and the interior metasedimentary rocks down dropped into the chamber, producing a chaotic assembly of broken rocks. These fragments sank into the top of the granitic magma chamber and were engulfed and surrounded by granite. This mixture now forms the shatter zone adjacent to the Cadillac Granite.

Driving Directions

On RT 102 about 0.5 miles south of Southwest Harbor, take RT 102A (left). At 3.0 miles, turn left into Seawall Picnic Area (Figure 35-1). Proceed to the loop at far right of parking area (Figure 35-2).

Figure 35-1. Map of parking lot and site of interest for this stop.

Figure 35-2. View of outcrops at Seawall picnic area (N 44º 14.206’ W 068º 18.184’). Aimee Dorais for scale.

Figure 35-3. Pumice fragments in the Cranberry Island Volcanics (N 44º 14.213’ W 068º 18.194’).

On the Outcrop

One can see many pumice fragments in this tuff (Figure 35-3). At shallow pressures, water exsolves from the magma and forms bubbles such as the myrolitic cavities seen in Figure 35-16. At times, the concentrations of bubbles becomes very high, forming a frothy cap to the magma chamber. Upon eruption, the frothy cap is disaggregated. These fragments are sponge-like and their high amounts of vesicles give the rock a very low density. The rock, called pumice, floats on water. Large pumice blocks are thought to have served as rafts for life forms to ride ocean currents and populate remote locations.

35b. Shatter Zone in Cadillac Mountain Granite

The 360 million year old Mount Cadillac Granite forms the highest peak in the park, Cadillac Mountain. It’s well worth the drive/hike to the summit to survey the park. Arriving just before sunrise is a popular activity as the summit of Cadillac Mountain is the first place in the United States to experience sunrise.

The intrusion of the Cadillac Mountain Granite was a violent event, incorporating fragments of the surrounding metasedimenary rocks that were broken into fragments during the collapse of the cauldera during the eruption of the Cranberry Island Volcanics. These fragments, called xenoliths after the Greek word for foreign, are highly concentrated at the margins of the Cadallac Granite, producing what is called a shatter zone between the granite and the surrounding metamorphic rocks. This zone is over a mile wide in some places and almost entirely surrounds the granite. This stop shows the visitor the many xenoliths or fragments of the surrounding metamorphic rocks that are now immersed in granite.

Driving Directions

Follow the Park Loop Road (Closed Dec-Apr) for 8.8 miles south from the visitor's center to Newport Cove (Figure 35-4). Park and walk to the east end of the beach (Figure 35-5).

Figure 35-4. Map of parking lot and the site of interest at this stop at Newport Cove.

Figure 35-5. Location of the site of interest along the eastern edge of the beach at Newport Cove.

Figure 35-6. The Cadillac Pluton is surrounded by a shatter zone, a zone of mixed rocks rich in fragments or xenoliths (foreign rock) broken from the rocks into which the granite intruded (N 44º 19.724' W 068º 10.808').

On the Outcrop

Figure 35-6 shows several different rock types that are floating in a matrix of Cadillac granite. Some fragments in the shatter zone are fragments of the metasedimenary rocks the pluton intruded, others are gabbro and diorite from an earlier pulse of magma. Still others are volcanic rock fragments as seen in the next photo (Figure 35-7). The shatter zone formed during intrusion of the granite as engulfed fragments of the surrounding rocks.

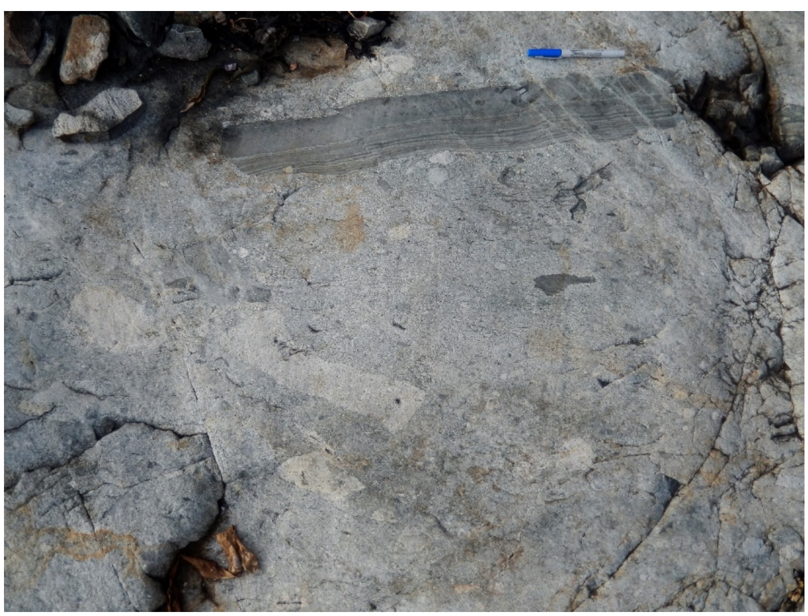


Figure 35-7. Light colored, volcanic rock fragments in the shatter zone.

35c. Schoodic Point

Driving Directions

Follow RT 1 east from Ellsworth ME for ~17 miles to RT 1 and RT 186 junction. Turn right on RT 186 for 7.0 miles to a large Acadia National Park sign. Turn right and proceed 11.5 miles and bear right. Continue to the parking lot at 12.1 miles (Figure 35-8).

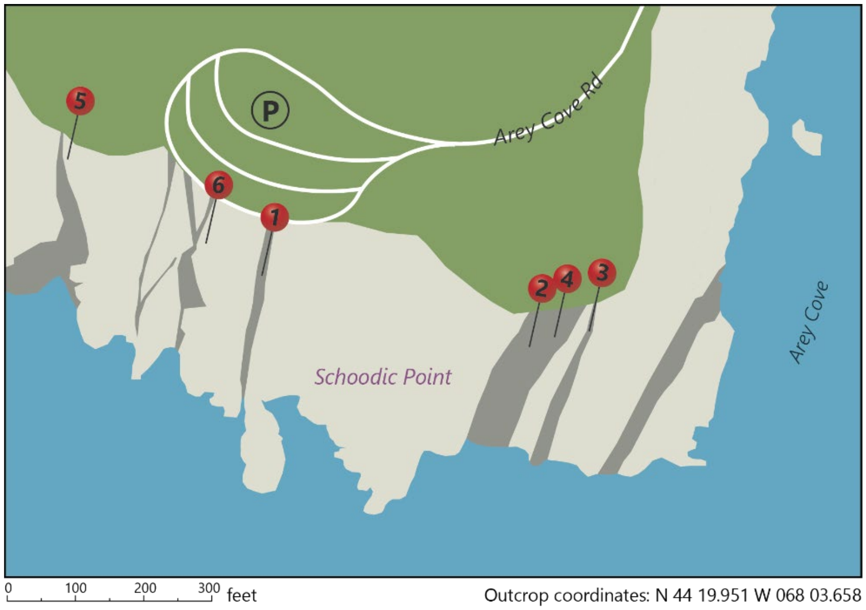


Figure 35-8. Map showing the location of the parking lot and sites to visit at this stop.

Figure 35-9. View to Mount Desert Isle across Frenchman's Bay. The rock in the foreground is the Devonian Gouldsboro Pluton.

Walking Directions

From the parking lot, proceed directly south to the rocks along the shore line (Figure 35-9).

Figure 35-10. Large Mesozoic dike cutting the granite at Schoodic Point at location 1 (N 44º 19.951' W 068º 03.658').

On the Outcrop

Location 1 (Figure 35-10): The chemical compositions of the basaltic dikes at Schoodic Point indicate that they correlate with dikes that are older than the Central Atlantic Magmatic Province dikes of Stops 22 and 42. The dikes at this location and those of Ogunquit (Stop 39) are from the very earliest rifting events of Pangea. They belong to the Coastal New England Province that was emplaced between 205 and 225 million years ago.

Location 2 (Figure 35-11): Some of the dikes at Schoodic Point are very large. This is the largest Coastal New England dike that I've seen, up to 60 feet thick. One can only image the massive amounts of basaltic magma that moved through this fracture.

Figure 35-11. Very wide basaltic dike of location 2 (N 44º 19.936’ W 068º 03.556’).

Figure 35-12. Dike tip of location 3 (N 44º 19.943’ W 068º 03.545’).

Location 3 (Figure 35-12): Most of the dikes that Schoodic Point represent fractures that have been enlarged and intruded by massive volumes of basaltic magma. These dikes represent an advanced stage of dike formation, far removed from the very first stage of magma intruding a fracture. One feature that makes Schoodic Point so interesting is that we can see the earliest stage of dike emplacement; i.e., we can examine the very tips of dikes as they just began to intrude the granite. Figure 35-12 shows the tip of a very small dike, usually called a dikelet. Note the rock to the right of the dike's tip is fractured and now filled with a greenish mineral called epidote. The epidote formed from the rock interacting with hot waters. If there had been a greater supply of basaltic magma to this dikelet, the magma could have forced the fracture to open further, producing a wider dike.

Figure 35-13. Two additional dikelets at Schoodic Point frozen in the act of following fractures extending to the right of the photo (N 44º 19.931' W 068º 03.548'). Ideally with perfect exposure, one could follow the dikelets to the left to a larger dike of basaltic rock.

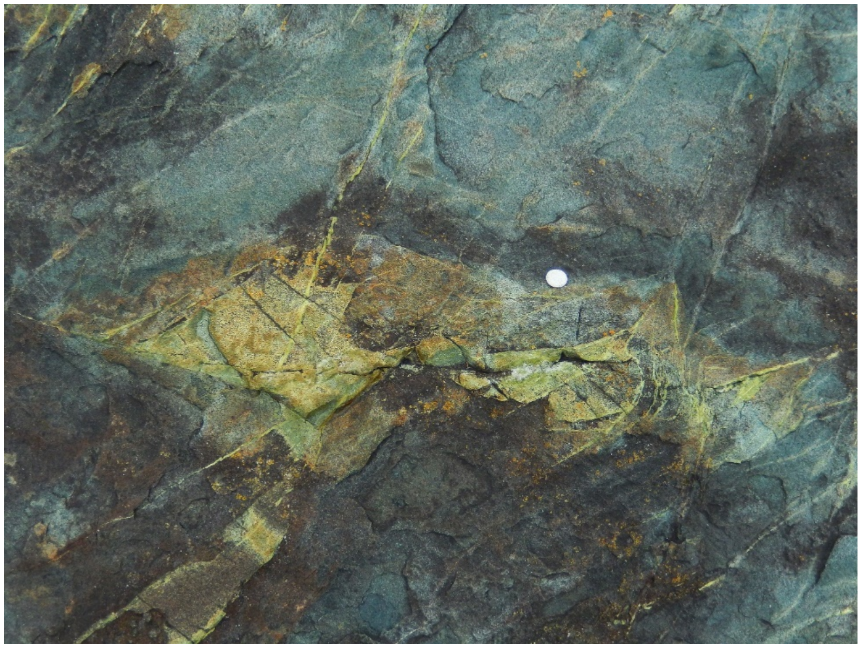


Figure 35-14. Hydrothermally altered dike at location 4 (N 44º 19.940’ W 068º 03.569’).

Location 4 (Figure 35-14): This dike shows extensive hydrothermal alteration and the formation of the mineral epidote ($Ca_2Al_2(Fe^{3+};Al)(SiO_4)(Si_2O_7)O(OH)$). Epidote is the pistachio green mineral. It did not form from the magma, but is a replacement mineral from interaction of hot water with the solidified dike.

Figure 35-15. Fault breccia in the Gouldsboro Pluton at location 5 (N 44º 19.976’ W 068º 03.723’).

Location 5 (Figure 35-15): The Devonian Gouldsboro Pluton was faulted prior to dike emplacement. Some of these show excellent examples of breccias (a rock with many angular fragments) along a fault zone, such as these in Figure 35-15. Note the angular blocks of granite that are engulfed in a finer-grained, dark matrix.

Figure 35-16. Myrolitic cavities in the Gouldsboro Pluton at location 6 (N 44° 19.958' W 068° 03.680').

Location 6 (Figure 35-16): I mentioned above that the granite at Schoodic Point was emplaced at shallow levels in the Earth's crust. Evidence for this shallow level is seen by the small cavities shown in Figure 35-16. These are myrolitic cavities that form from bubbles in the magma. Water dissolves in granitic magmas, the higher the pressure on the magma, the more water will dissolve. But like soda that forms bubbles when the can is opened (i.e., pressure is decreased allowing the CO_2 to exsolve from the soda), so do bubbles form in a magma as it ascends, decreasing the pressure. The rock at this location looks pock marked from all the water bubbles. Strictly speaking, the bubbles didn't contain liquid water; at magmatic temperatures the water was in a vapor state. If you look closely at the cavities, you'll note that some have crystals that grew into them. Some of these crystals grew from the magma, others from dissolved components in the vapor. Magmatic vapors are able to dissolve considerable silica and other elements that precipitate on cooling.

36. Deer Isle Granite, Flye Point, ME

The Deer Isle Granite is one of the most beautiful granites in all New England. It contains abundant rapakivi feldspars, rounded potassium feldspar (Kspar) grains that are mantled by plagioclase feldspar. This type of granite was originally studied in Finland where the Finish word rapakivi was used to describe the crumbly nature of the rock in those outcrops. Fortunately for us, the rapakivi granite at Flye Point is not crumbly at all, superbly fresh outcrops are abundant along the coastline.

The Deer Isle Granite is one of many plutons along the Maine coast, in fact, their abundance prompted geologists to group them as the Coastal Maine Magmatic Province. Other plutons in this guide of the Coastal Maine Magmatic Province include the rocks at Acadia National Park (Stop 35), and Vinalhaven Island (Stop 37) (Figure MS-2).

Plutons of the Coastal Maine Magmatic Province offer some of the best locations to examine the effects of magma mixing, particularly mafic magmas interacting with granitic magmas. Sometimes the evidence for mixing is overwhelmingly strong as at Vinalhaven Island; other plutons are more discrete, masking their history of magma mixing. Such is the case at Deer Isle. Most geologists think that rapakivi texture is the result of injection of a more mafic magma into the granitic pluton, heating the granitic magma to the point where Kspar is not stable, transforming rectangular Kspar crystals into the rounded, ovoid Kpars as the crystals started to dissolve. Since plagioclase tends to crystallize at a higher temperature than Kspar, the heating from the mafic magma does not destabilize plagioclase crystallization. Plagioclase then grows on the Kspar ovoids, forming the distinctive rapakivi texture.

Driving Directions

Just east of Bucksport, ME, follow RT 1 east to RT 15 South (Front Ridge Road). Follow RT 15 south to Blue Hill. At Blue Hill, turn right on RT 172. In 0.6 miles, turn left on RT 175/172. Follow for 2.3 miles to junction where RTs 175 and 172 split, take a left on RT 175 (Falls Bridge Road). In 8.1 miles at a ~ 90 curve on RT 175, take a left on Flye Point Road. The road is not well marked, so watch for signs to the Lookout Inn. Follow Flye Point road to the Lookout Inn and proceed for 0.3 miles past the Inn to a grassy field and park at the end of the dirt road just north of the gazebo (Figure 36-1).

Figure 36-1 showing the location of the outcrops along the shore at Flye Point.

Walking Directions

Excellent outcrops are located along the shore line at N 44º 15.662' W 068º 31.161'(Figure 36-2).

Figure 36-2 Fresh outcrop of the Deer Isle Granite at Flye Point.

Figure 36-3. Pink Kspar crystals rimmed by white plagioclase, forming rapakivi texture in the Deer Isle Granite. Darker gray minerals are quartz grains; black minerals are biotite and amphibole.

On the Outcrop

The pinkish crystals are Kspar, the white are plagioclase (Figure 36-3). Note that many of the Kspar crystals are rounded. They typically grow as rectangular crystals but become rounded when they are heated beyond their stability temperatures. Imagine immersing a sugar cube in hot water. Eventually the cube will lose its corners and become rounded. In the same manner, the originally rectangular Kspar crystals became rounded and then coated by plagioclase feldspar to form the white rims.

Figure 36-4. Mafic enclave or inclusion in the Deer Isle Granite.

The Deer Isle Granite at Flye Point contains many fragments of darker, finer grained rock (Figure 36-4). The dark color is from fine-grained biotite and/or amphibole, similar to enclaves seen in plutons at other locations (Stop 16, 17b, 23, and 37). These dark rocks are called mafic enclaves or mafic inclusions and represent blobs of more mafic magma that were injected into the Deer Isle Granite and mixed with it. The abundant feldspar crystals in the enclave came from the Deer Isle granitic magma as the two magma types were stirred together. This enclave is probably a fragment of a mafic dike that injected into the Deer Isle magma chamber that may have been the heat source to heat the granite and round the Kspar feldspar crystals.

37. Vinalhaven Island, ME

The residents of Vinalhaven Island might not realize that their island is world famous to geologists. Perhaps the term “world famous” is a bit of an overstatement because the number of geologist in the world isn’t large. But among geologists who study igneous rocks, Vinalhaven is considered a classic. Why would geologists from overseas be interested in a small island off the coast of Maine? Because Vinalhaven has some of the best examples of the interaction between mafic and felsic magmas anywhere.

The Vinalhaven Island Complex is one of many plutons that constitute the Coastal Maine Magmatic Province. These magmas resulted from subduction under the Ganderian microcontinent as it approached and collided with Laurentia during the Salinic Orogeny about 430 million years ago. The province is famous because of the excellent exposures along the coast line that show mixing of mafic and felsic magmas. Vinalhaven is the best of the best, well worth the cost of the ferry to see such spectacular rocks.

Granite has been quarried on Vinalhaven Island since 1826, and for the next century, Vinalhaven became one of Maine's largest quarrying centers. You’ll note several abandoned old quarries on your way to Lane Island Preserve. Pinkish-gray Vinalhaven granite has been used in the construction of the State Department Building in Washington, New York City's Brooklyn Bridge, and the Union Mutual Life Insurance Building in Boston, the Board of Trade in Chicago, and the Washington Monument.

Driving Directions

Drive to Rockland, ME and park at the ferry at 517A Main Street, RT 1. At the time of writing, the fare for a car to be ferried to Vinalhaven Island is $49. The passenger fare is $17.50. The sites described below are within walking distance (about 1.5 miles to the outcrops) and a car is not needed for most visitors. Please note that passengers must have return reservations to return to Rockland or you could be stuck overnight on Vinalhaven.

Figure 37-1. View of Rockland Breakwater Lighthouse leaving Rockland Harbor with Mt Waldo, underlain by a Devonian granite, in the background.

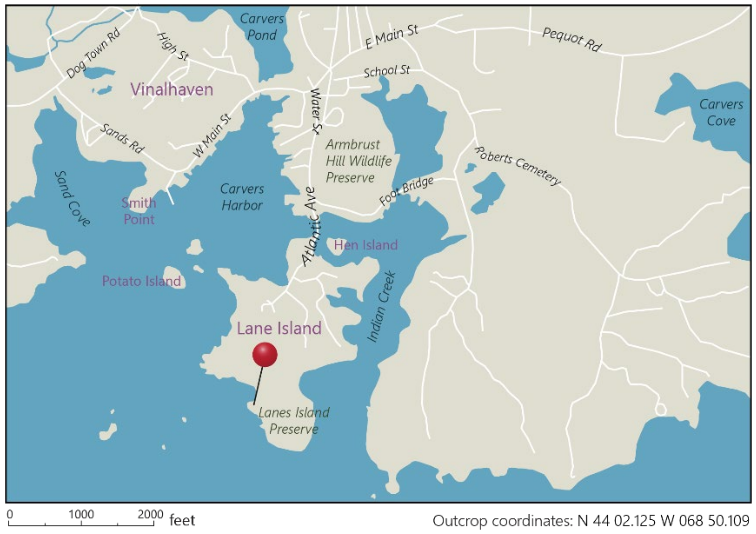


Figure 37-2. Map of a portion of Vinallhaven Island showing the harbor and roads to Lane Island. Outcrops for this stop are along the southwest shore of the island.

Walking Directions

From the ferry dock, walk east along Main Street for 0.48 miles and turn south on Water Street (Figure 37-2). After 0.1 miles, continue straight on Atlantic Ave where you'll cross a bridge to Lane Island in ¼ of a mile. At 0.15 miles from the south end of the bridge, turn left on Lane Island Preserve Road. Continue past parking area at 0.13 miles down the dirt road (park here if you drove a car) and proceed along the path at the south end of the parking area. After 350 feet, you'll arrive at a large, grassy field. Follow the path that lead due west. You'll see a cemetery marker on the left. A trail starts at a wooden sign-in-box at N 44º 02.237' W 068º 49.959'. Follow trail to shore line (Figure 37-3).

Figure 37-3. Photograph of a shaded outcrop with abundant basaltic pillows (N 44º 02.125' W 068º 50.109').

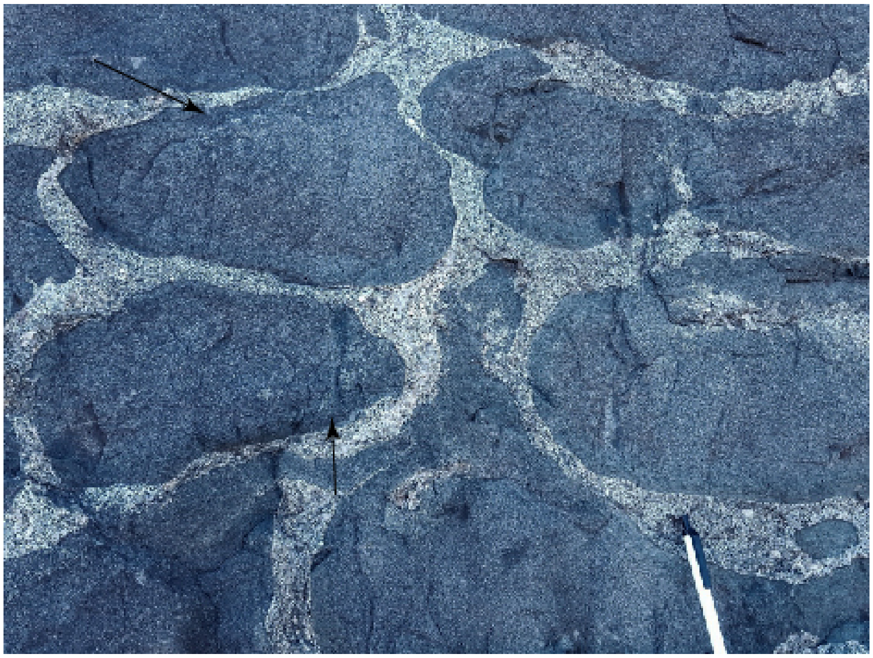


Figure 37-4. Photograph of pillow basalts in the granitic pluton separated by thin rinds of granite (N 44º 02.125’ W 068º 50.109’). Arrow points to a chilled margin on one of the pillows.

Figure 37-5. Pillows in the Vinalhaven Granite (N 44º 02.098’ W 068º 50.091’).

On the Outcrop

Basaltic lavas that erupt under water form features called pillows. The large temperature contrast between the lava and the water causes the lava to quench, especially at the interface where the lava solidifies to a fine-grained crust. As more lava flows inside the solidified rind, the pressure can be large enough to rupture the rind and the lava flows out of the rupture in a tube-like shape, much like toothpaste is squeezed from its tube. Repeating this process produces a pile of interconnected lobate shapes that are pillow shaped in cross section. As the pillows pile up, they tend to be flat on the bottom and rounded at the top, giving an up direction indicator.

What is so interesting here on Vinalhaven Island is that basaltic magmas that injected into a granitic magma chamber formed pillows too (Figures 37-4 and 37-5). This spectacular outcrop shows many pillows with intervening granitic rocks filling the space between the pillows. The temperature contrast between the basaltic and granitic magmas is not as strong as that between basaltic lava and water, perhaps only a few hundred degrees Celsius, but nonetheless, pillows of the same shape are formed. The quenching of the basaltic magma forms finer grained pillows whereas the granitic magma stayed molten and crystallized to form minerals of larger grain size. Sometimes, these pillows even preserve up direction indicators as do their marine cousins, preserving rounded tops and flat bottoms, as well as lateral flow directions when pillows branch and split.

Figure 37-6. Pillow showing amphibole-rich rind resulting from absorption of water into the dryer basaltic magma. Note the finer grain size along the chilled margin of the pillow (e.g., under the quarter).

As noted at Stop 35c, granitic magmas contain dissolved water in the same manner as CO_2 is dissolved in soda. The basaltic magma that injected into the granitic pluton was dryer and hotter than the granitic magma. The dark rind on the pillow in Figure 37-6 was formed as water from the granitic magma migrated into the pillow while the granite was still molten, transforming pyroxene in the pillow to amphibole. Water constitutes a few percent of the amphibole structure. The presence of amphibole gives the pillow the darker color along its edge. Water didn't migrate into the center of the pillow, hence, the rock still contains its original pyroxene.

Figure 37-7. Pillow fragments from disruption of quenched basaltic magma.

Because the basaltic magma was hotter than the granitic magma by a few hundred degrees Celsius, the basaltic cools against the granite and the granite heats up against the basalt. Which has the greater influence depends on the volume of each magma type; where the granitic magma dominates over small amounts of basalt, the basalt is quickly quenched to fine-grained rock such as that seen at Stops 16, 17b, 23, and 36. The basalt then is much more viscous and in some cases where the temperature contrast is high, the basalt solidifies and fragments into smaller pieces. Figure 37-7 shows many small fragments of basaltic rock that have sharp edges. Some are elongated, representing busted pillows.

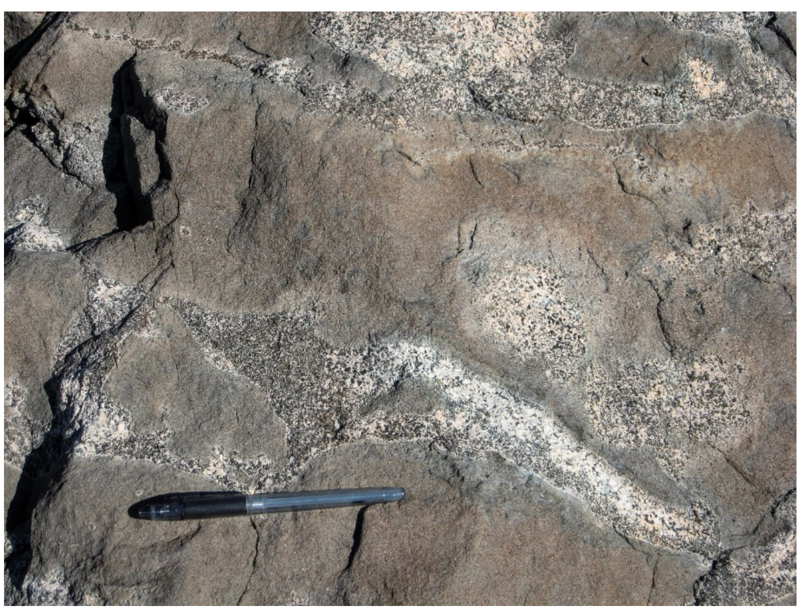


Figure 37-8. Hybridization of magmas produced rocks of intermediate composition between the pillows.

When the basalt isn't totally quenched against the granitic liquid, remaining a mush of liquid and crystals, the two liquids can mix. Note the abundance of darker minerals in the granitic zones between the pillows in Figure 37-8. These zones are hybrids of the basalt and granitic magmas and have compositions in between the two.

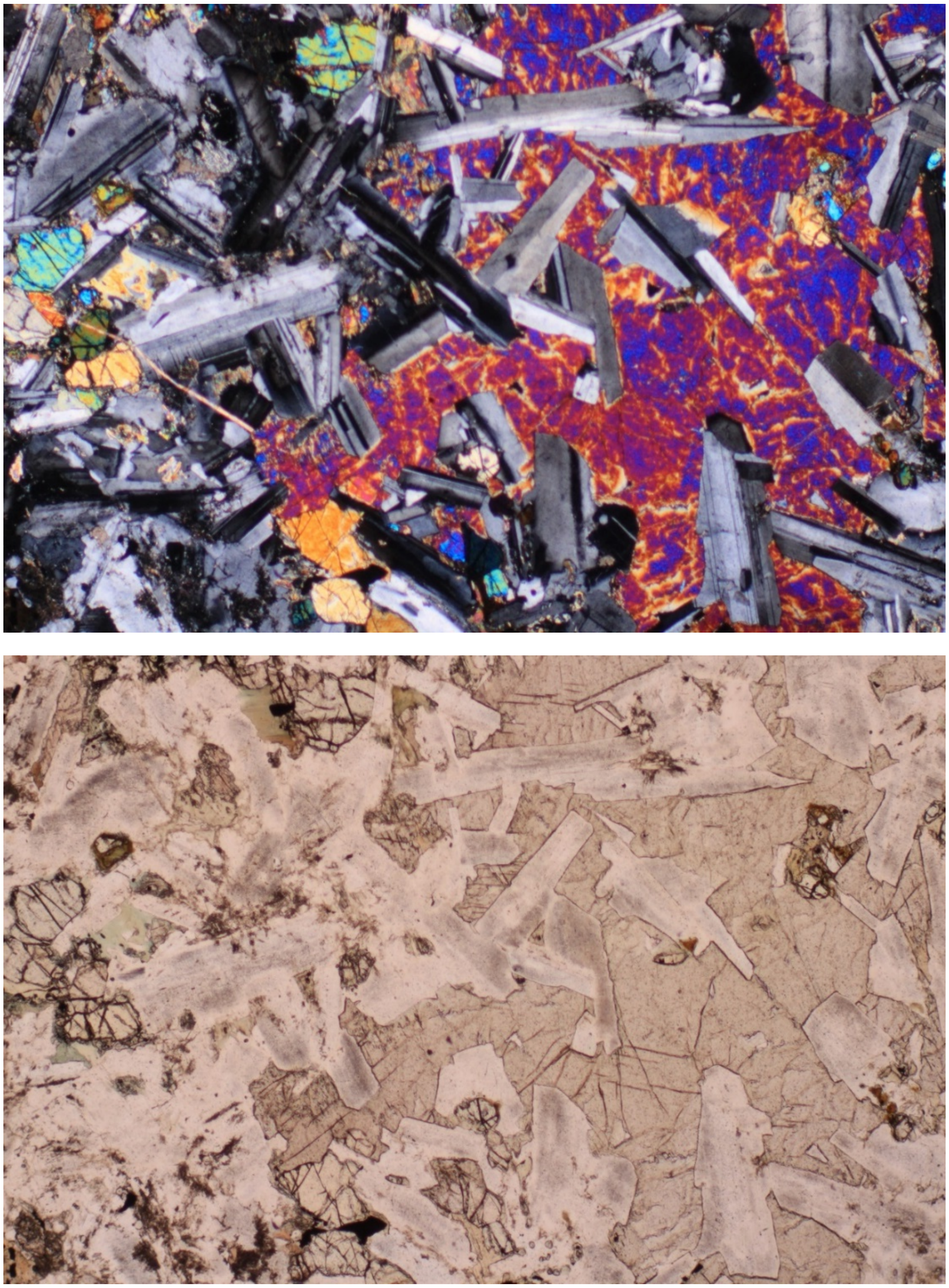


Figure 37-9. Crossed polarized light (upper photo) and plane polarized light (lower photo) photomicrographs of a gabbroic sample of the Vinalhaven Complex. The rock consists of plagioclase feldspar (gray and white crystals), pyroxene (orange to blue mineral with plagioclase inclusions, and olivine (small greenish mineral in the left portion of the upper photo).
Field of view is 1 mm.

38. Bucksport Formation, Pemaquid Point, South Bristol, ME

For sheer beauty, it's hard to beat Pemaquid Point. It must be a pretty location because when the US mint issued quarters commemorative of each state, Maine's was imprinted with an image of the Pemaquid Point lighthouse. Not only does this location have a classic Maine lighthouse, it has wonderful outcrops of metamorphic and granitic rocks. The metasediments were deposited as muddy sands and silts in the Frederickton Trough about 430 million years ago (Figure MS-1). They were metamorphosed to schists and gneisses and intruded by granites and their pegmatites during the Acadian Orogeny. But the granites weren't intruding into a static body of metasediments. The entire region was being deformed during movement along the Norembega Faults that is about 20 miles north of Pemaquid. This fault system rivals the San Andreas Fault in length, with coastal Maine sliding to the southwest compared to inland Maine. The metasedimentary rocks at Pemaquid Point were tightly folded in the process. Granitic dikes from the 367 million year old Waldoboro Pluton were intruded during and after this deformation. Some of the granitic sheets are deformed, pulled apart and rotated into the direction of shearing.

Driving Directions

Take RT 1 to Newcastle, ME. Continue through Newcastle on RT 1 to junction with RT 129. Turn south (right) on RT 129 and follow for 3.7 miles to the junction with RT 130. Turn left on RT 130 and follow this road for 15.5 miles to Pemaquid Point Lighthouse Park (Figure 38-1).

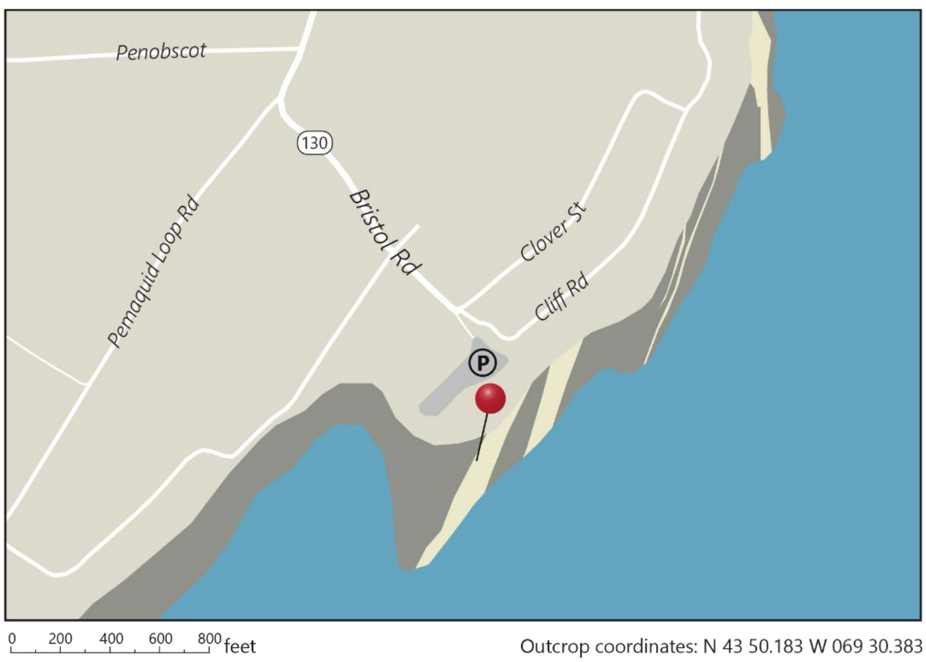


Figure 38-1. Map of Pemaquit Point showing the parking lot, light house and outcrop locations.

Walking Directions

From the lighthouse, walk down to the excellent exposures of outcrop (Figure 38-2).

Figure 38-2 (left). A large dike of granitic rock intruded the metasediments (N 43º 50.183’ W 069º 30.383’). Note the orientation of the dike is parallel to the layering of the metasediments which is near vertical.

Figure 38-3 (right). Pegmatitic dikes were broken and pulled apart during deformation associated with movement along the Norembega Fault.

Figure 38.4 (left). Some pegmatitic dikes were stretched and pulled apart after solidification.

Figure 38-5 (right). Blocks of metasediment were incorporated into the granitic dikes (N 43º 50.232’ W 069º 30.306’). Some of these blocks display folding as seen in the block in the center of the photo.

Figure 38-6. Some of the granitic dikes are very coarse-grained rocks called pegmatites. The light brownish white crystals are feldspar.

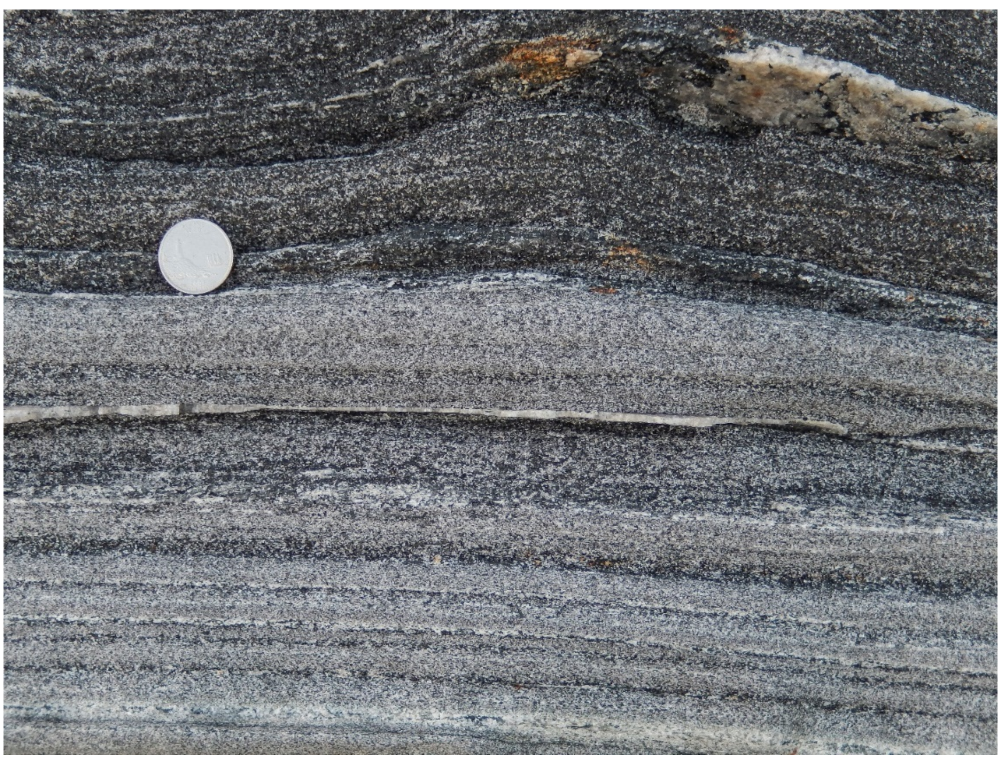


Figure 38-7. The metasediment is heterogeneous, with the different colors of the layers determined by the original composition of the sediment.

On the Outcrop

The darkest layer at the top of the photo of Figure 38-7 contains more biotite than the lighter gray layers that are more feldspar and quartz rich. A thin quartz vein cuts the rock across the middle of the photo. Some thin bands in the metamorphosed sediments are greenish in color. These mark deposits of sediment that were richer in calcium than the dark layers. Upon metamorphism, the calcium combined with other elements for form a pyroxene called diopside.

The Acadian Orogeny transformed these sediments to metamorphic rocks that no longer maintain their original, horizontal orientation. The rocks below the light house of Figure 38-2 have been tilted to near vertical position, and as seen in Figure 38-8, the rocks have also been folded. During this deformational event, granitic magmas were emplaced. Some of these magmas were solidified and were subsequently deformed along with the metasediments they intruded. Figures 38-3 and 38-4 show felsic rocks that were once sheet-like dikes; now they are broken up lenses because they were more rigid that the metasediments that were more susceptible to flow during deformation.

Figure 38-8. Folded Bucksport Formation metasediments.

39. Kittery Formation, Merrimack Trough, Marginal Way, Ogunquit, ME

The Kittery Formation represents sediments shed from the approaching Ganderian microcontinent during the Silurian Period. Rivers eroded the sediment from that microcontinent, bringing them to the Merribuckfred Basin which lay between the microcontinent and Falmouth-Brunswick/Casco Bay island arc (Figure MS-1). The sediments represent turbidity deposits, i.e., density currents that avalanched off the microcontinental slope down into the deep water basin.

Turbidity currents are fast-flowing subaqueous debris flows that carry high amounts of sediment in suspension from the edge of the continental shelf down the continental slope to the ocean floor (See Figure 29-1). These sediment-rich avalanches, probably triggered by earthquakes along the then tectonically active margin, cascaded down slopes under the influence of gravity because mixed sediment-water suspensions have higher densities than sediment-free water. Mud, sand, and water form a dense slurry in the currents. Observations of modern turbidity currents indicate that they move at speeds of over 35 miles per hour, traveling with tremendous erosive power, scouring channels and canyons into the underlying sediments. The turbidity currents spread out horizontally along the floor of the basin forming submarine fans, and as the current slows down, the denser, more coarse-grained sediments are deposited first with the finer-grained sediments and muds still in suspension. When water velocities are slow enough, even the mud will settle out of the water column. Deposits from turbidity currents are therefore layered with coarse-grained sediments at the bottom and clay-rich sediments at the top of each layer. This coarse- to fine-grained characteristic enables us to determine the upward direction of the Kittery Formation and other turbidity deposits, i.e., which layers are oldest and which are youngest. One might think that because the law of original horizontality requires the oldest layers to be at the bottom of a sedimentary sequence, hence the younging direction should be easily determined. However, in orogenic belts, rocks are tilted and folded and the younger rocks may not necessarily be higher in an outcrop if the sequence has been tectonically flipped. At this stop, we will see folded Kittery Formation rocks that have been inverted where the lower limb of the fold shows upside-down grade bedding.

The Kittery Formation, like other trough sediments described in this guide such as the Appleton Ridge Formation, were deformed during the Devonian Period as the Avalonian microcontinent collided with Laurentia during the Acadian Orogeny, though the degree of metamorphism of these rocks is much lower than some of the metasediments of New Hampshire (e.g., Stop 19). Here we can examine the coarse- and fine-grained sediments derived from turbitity currents (also see the Carrabassett Formation of Stop 29), folding of these sedimentary layers from Acadian collision, and much later intrusion of basaltic and granitic dikes related to initial stages of continental breakup. Many basaltic dikes are present here, not as thick as the dikes at Schoodic Point at Acadia National Park (Stop 35c), but interesting in their own right.

The Kittery Formation is part of an enigmatic package of rocks called the Merrimack Trough, a portion of the Merribuckfred basin shown in Figure MS-1. It is the smallest tectonic fragment in New England and is generally accepted that the Merrimack Trough contains sediments that were derived from an outboard, non-North American terrane. Most geologists think the sediments were shed from Ganderain source and are therefore related to other coastal blocks along the coast of Maine and Maritime Canada.

This stop occurs along the Marginal Way, a very popular public footpath offering outstanding views of the Maine coastline in Ogunquit.

Driving Directions

Take RT 1 to Ogunquit, ME. In downtown Ogunquit, turn east on Shore Road. Follow Shore Road for 0.4 miles to Israel Head Road on the left. Talk Israel Head Road for 0.3 miles, park on the left near the miniature lighthouse (Figure 39-1). During the summer months, these few parking spaces will probably be occupied. If so, return to Shore Road and turn left and drive to Perkins Cove Road. There are several pay lots along Perkins Cove. Please note that there may be a time limit for parking in some of these lots.

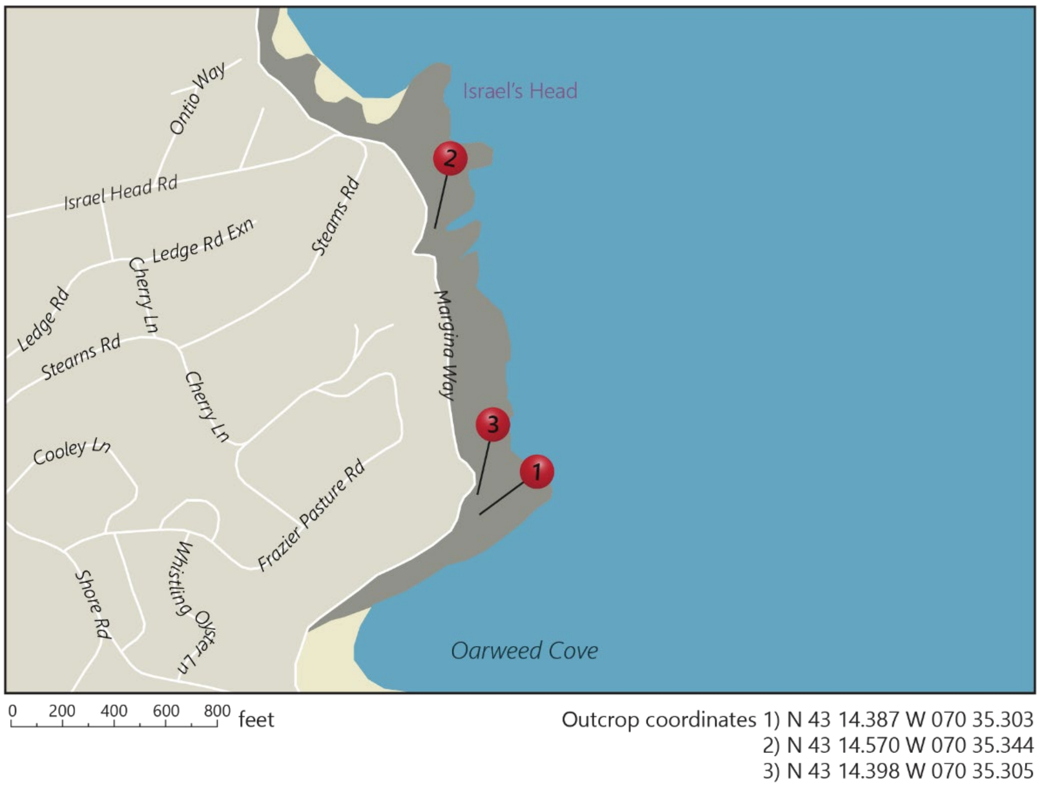


Figure 39-1. Map of the Marginal Way path at Ogunquit, ME. Circles with numbers indicate sites of interest along the Marginal Way.

Walking Directions

If you parked at the mini lighthouse on Israel Head Road, follow the Marginal Way path south. If you park at Perkin's Cove, follow the footpath to the north of the parking lot.

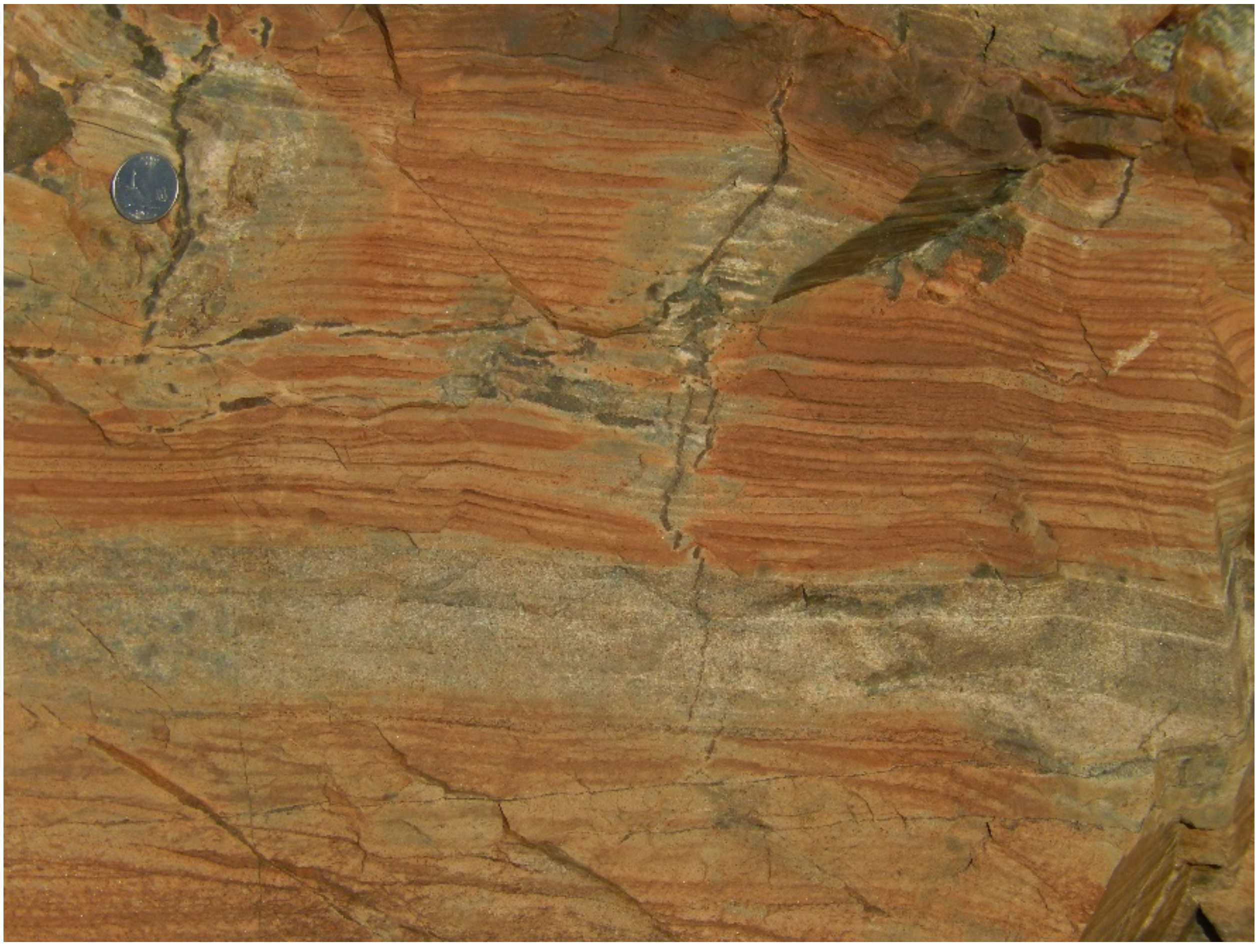


Figure 39-2. Turbidity current deposits of the Kittery Formation.

On the Outcrop

Figure 39-2 shows a sandy layer from a turbidity current in the middle of the photo. This lighter tan layer is mostly quartz and because it has been metamorphosed, the rock is called quartzite. This is the denser, coarser-grained portion of the deposit that fines upward. The finer-grained, darker brown, thinly laminated layer under the quarter is the more clay-rich portion of the current, now a metamorphic rock called phyllite. The coarser-grained quartzite and the overlying phyllite constitute one turbidity deposit. See description of Stop 29 for additional examples of turbidity current deposits.

Figure 39-3. Overturned fold in the Kittery Formation (N 43º 14.387’ W 070º 35.303’).

This outcrop shows the Kittery Formation that was folded during the Acadian Orogeny. The beds at the top of the outcrop (Figure 39-3) maintain their original upward orientation with the younger beds at the top. The lower portion of the outcrop is the opposite; the youngest beds are at the base of the outcrop, similar to the schematic diagram of Figure 38-4.

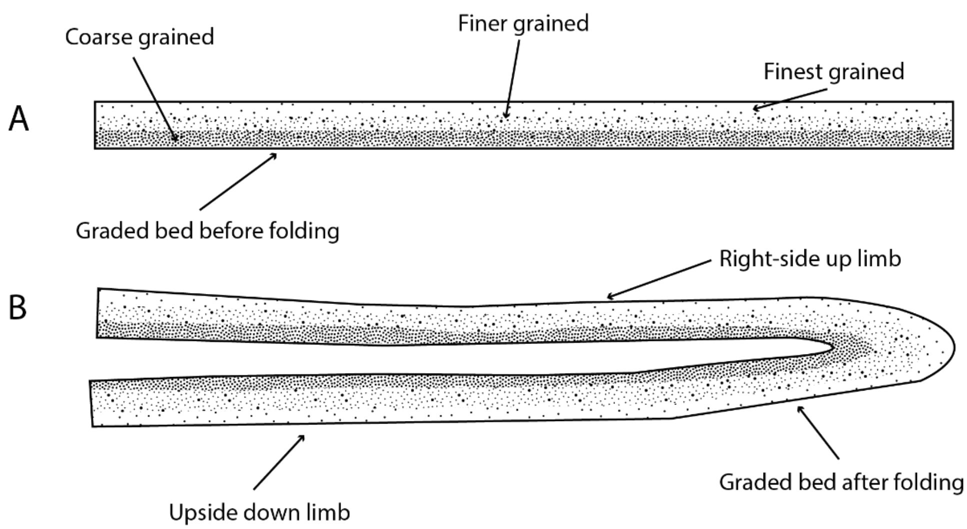


Figure 39-4. A) Schematic diagram showing a graded bed with coarse- to fine-grain particle size from bottom to top. B) Folded bed showing the right-side up limb where the grain size gradient is normal versus the upside down limb where the grain size ranges from fine to coarse from bottom to top of the bed (After Hussey, 2000).

Figure 39-5. Axial plane cleavage in the Kittery Formation (N 43º 14.570’ W 070º 35.344’).

Here you can easily see two different types of sediment that form layers extending from the lower left to upper right in Figure 39-5. The lower layer is quartzite whereas the upper layer represents finer grained phyllite of a single turbidity deposit. The upper left portion of the outcrop shows thinner bedding in the phyllite. But the feature that catches one’s eye is the sub-horizontal lines that seem to cut across the phyllite, giving the outcrop a layered effect. These “layers” are not bedding, but result from folding of the sediments. The rocks between the fold limbs were compressed during folding, forming axial plane cleavage. The phyllite contains abundant platy minerals like the micas biotite and muscovite that adjust to that deformation, aligning themselves perpendicular to the stress field squeezing the rocks into folds to form cleavage. The quartzite seems oblivious to the deformation, not because it didn’t experience the same stress as the phyllite, but because more rounded grains of quartz -rich nature of the rock are not as reactive and lack sheet-like micas.

Figure 39-6. Basaltic dike cutting a felsic dike in the Kittery Formation (N 43º 14.398’ W 070º 35.305’). The Kittery Formation is shown in the upper right portion of the photo.

Here we see a granitic dike that perhaps emanated from the nearby Biddeford pluton, cut by a younger basaltic dike (Figure 39-6). The basalt is from the same melting event that produced basaltic dikes at Stop 35c and is related to the early stage of continental rifting.

Figure 39-7. Basaltic dike with vesicles in the Kittery Formation (N 43º 14.367' W 070º 35.340').

Like the vapors that formed myrolitic cavities in the Gouldsboro Granite at Schoodic Point (Stop 35c), basaltic magmas can also exsolve vapors at low pressures. Figure 39-7 shows a small dike with abundant vesicles aligned in the center of the dike. None seem to be present along the dike's margin. As the basaltic magma intruded along the fracture, the outer margins of the dike were quenched against the relatively cold Kittery Formation. The dike's margin just above the quarter is finer grained than the interior of the dike. As the magma continued to flow, the bubbles and sometimes mineral crystals are concentrated where magma flow was highest. In some dikes, these vesicles can be filled with minerals that crystallize from vapors.

Figure 39-8. Cross-cutting dike in the Kittery Formation (N 43º 14.367’ W 070º 35.340’).

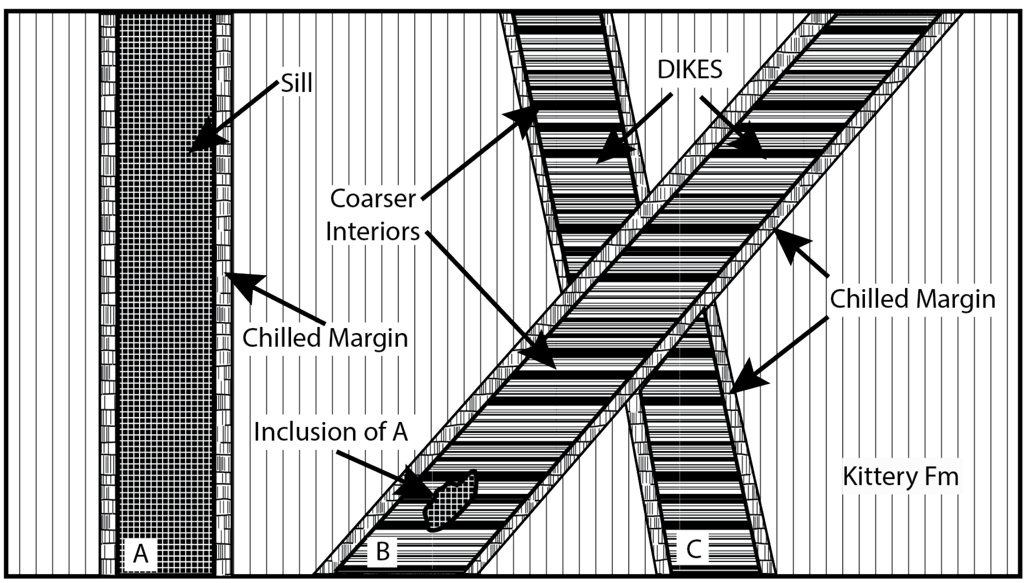


Figure 39-9. Schematic depiction of a cross-cutting dike showing similar features to those seen in Figure 39-8 (After Hussey, 2000). The vertical lines in the Kittery Formation represent bedding. A sill (A) intrudes parallel to the bedding in the host metasediments whereas dikes (B and C) cut across it. Dike B is younger than dike C.

At this location, you’ll see cross cutting dikes with an older dike having been cut by a younger one (Figure 39-8). Look to see if you can find chilled margins in the younger dike where the basaltic magma chilled against the older, cold basaltic dike.

40. Kittery Formation, Two Lights State Park, Cape Elizabeth, ME

The rocks at Two Lights State Park belong to the Kittery Formation, the same formation seen at Ogunquit at Stop 39. The sediments were deposited during the Silurian Period, about 430 million years ago, into an ocean basin that lay between the Laurentian continent and the approaching microcontinent of Ganderia (Figure MS-1). Like the rocks at Ogunquit, these also show rhythmically layered quartzite and phyllite beds of variable thicknesses. What makes these outcrops interesting is the rather unusual grain or cleavage of the rock that was formed when the rocks were compressed and deformed during the Acadian Orogeny. An additional point of interest is the presence of a fault that produced sufficient friction to partially melt small amounts of the Kittery Formation to produce pseudotachylite similar to that seen at Jones Corner of Stop 33.

Driving Directions

Take I-95 to Portland, ME. Then take I-295 towards South Portland at Exit 4. Cross the Veterans Memorial Bridge and follow W Commercial Street towards the Cosco Bay Bridge. Turn left on the Casco Bay bridge and follow signs to RT 77 South. At 6.4 miles, turn right to RT 77 Cape Elizabeth (Ocean Street). At 11.0 miles, turn left on Two Lights Road. At 12.0 miles, enter Two Lights State Park on right (Figure 40-1).

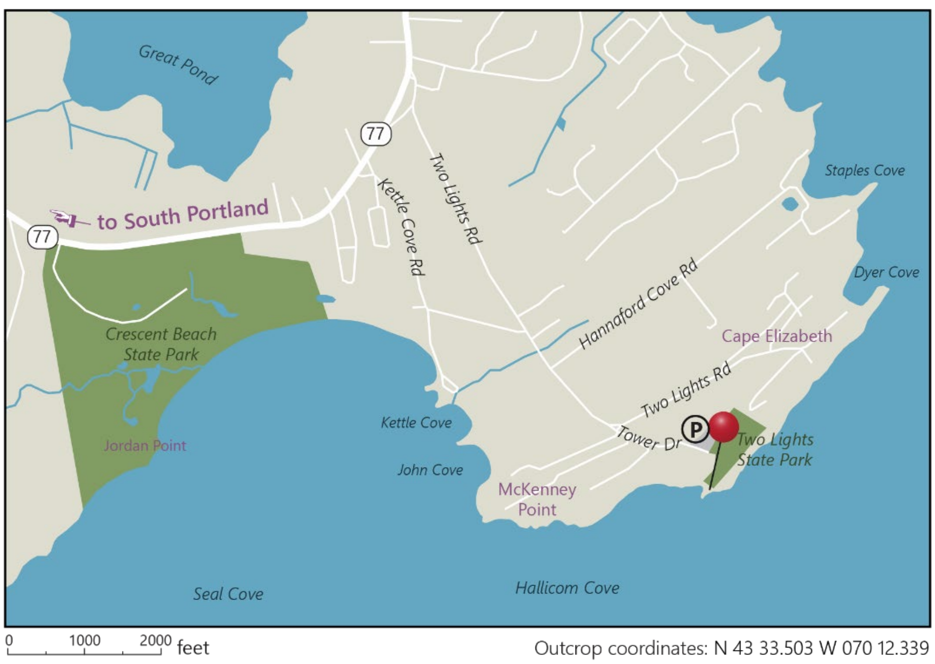


Figure 40-1. Map of Two Lights State Park at Cape Elizabeth, ME.

Walking Directions

From the parking lot, follow the main path past the fortifications to a small cliff. Descend the cliff to the right.

Figure 40-2 showing outcrops along shore at Two Lights State Park. Location of subsequent photos is the opposite direction from this view.

Figure 40-3. Well defined cleavage is displayed at Two Lights State Park.

On the Outcrop

During metamorphism and deformation, new minerals grow in a preferred alignment forming a fabric to the rock. Geologists call this fabric cleavage. The rocks at Two Lights show some of the most unusual cleavage because it gives the rock the appearance of petrified wood, especially the rocks under the GPS devise of Figure 40-3.

Other features to note here are the different types of bedding. Figure 40-4 shows a thick bed of quartzite among finer-grained phyllite layers. Along the phyllite layer is a folded and broken quartz vein. Quartz veins form from hot fluids that contain dissolved silica. As these fluids cool, the silica precipitates to form quartz. The rock was deformed after formation of the quartz vein, folding and breaking it. Figure 40-5 shows a quartz vein that was stretched and broken during deformation of the metasediments. The phyllite is more mobile, flowing easier than the quartz vein, pulling apart the quartz vein in the process.

Figure 40-4. This photo shows a thick bed of folded quartzite with a folded, thinner band of phyllite. Stretched and broken quartz veins are also present.

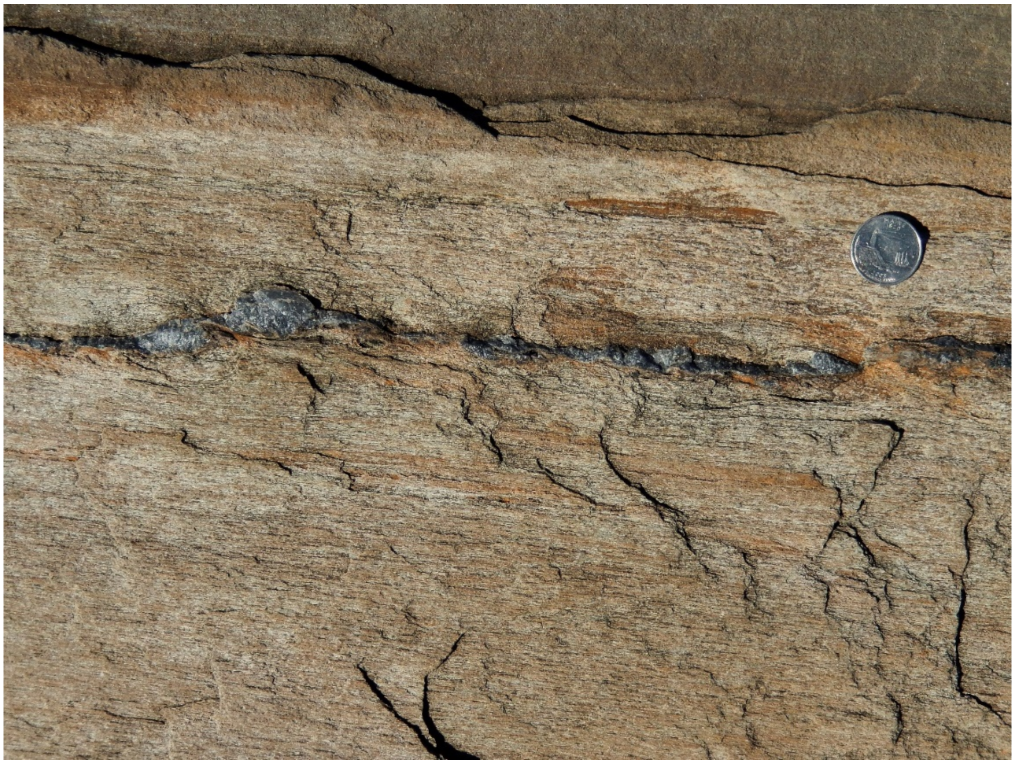


Figure 40-5. Photo showing a stretched and pulled apart quartz vein in the phyllite (N 43º 33.504' W 070º 12.333').

Figure 40-6. Fault cutting the Kittery Formation. A careful examination of the fault will reveal the presence of pseudotachylite (N 43º 33.503' W 070º 12.339').

Look carefully at the base of the small cliff shown in Figure 40-6. A vertical fault separates the sun-lit block to the left from the more shaded block to the right. This is a right lateral fault, meaning the block to the right moved toward the viewer as the block to the left move away.

The close-up photo of Figure 40-7 shows a finer-grained, darker material occurring along the fault surface. This dark rock is pseudotachylite. Pseudotachylite forms from frictional melting as the two blocks slide past each other under considerable pressure. The melt rapidly crystallized to glass, much of which is now devitrified to fine-grained minerals. This band of pseudotachylite represents a seismic event that generated an earthquake produced along the Norembaga Fault System (also see Stop 33) as the outboard block slid to the southwest with respect to the inboard block.

Figure 40-7. Pseudotachylite along the fault plane (N 43º 33.504' W 070º 12.333').

41. Cape Neddick Gabbro, York, ME

The Nubble Lighthouse on Nubble Island lies about 100 yards off Cape Neddick Point. The lighthouse has been in operation since 1879. Its accessibility makes it one of the more visited and photographed lighthouses on the Maine coast. Some of these photographs ended up in an unusual place with potentially, unusual viewers. The spacecraft Voyager carries photographs of some of the Earth's most significant landmarks and man-made structures. Included in the collection is a photo of the Nubble lighthouse. Interested extraterrestrials will know where to visit, perhaps sampling some of the excellent seafood offered in the region.

For Earthbound visitors, the 116 million year old, Cretaceous Cape Neddick Gabbro is also interesting. Gabbro is an igneous rock that consists of plagioclase feldspar and pyroxene $\pm$ olivine or amphibole. Many gabbroic plutons are layered with the layering having formed from a variety of processes in the magma chamber. The simplest explanation for igneous layering is that crystal laden currents swept across the floor of the magma chamber, depositing crystals like sediments in successive layers. At Sohier Park, one can examine the faint layering in the pluton. Amphibole-bearing pegmatites are also easily observed here.

Driving Directions

From York Beach, ME, follow RT 1A (Long Beach Road) north to Nubble Road. Turn right on Nubble Road for 0.9 miles ot Sohier Park Road. Park at end of Sohier Road (Figure 41-1).

Figure 41-1. Map of the Sohier Park area and the Nubble Lighthouse, York, ME.

Figure 41-2. Photo of the Nubble Lighthouse from Sohier Park.

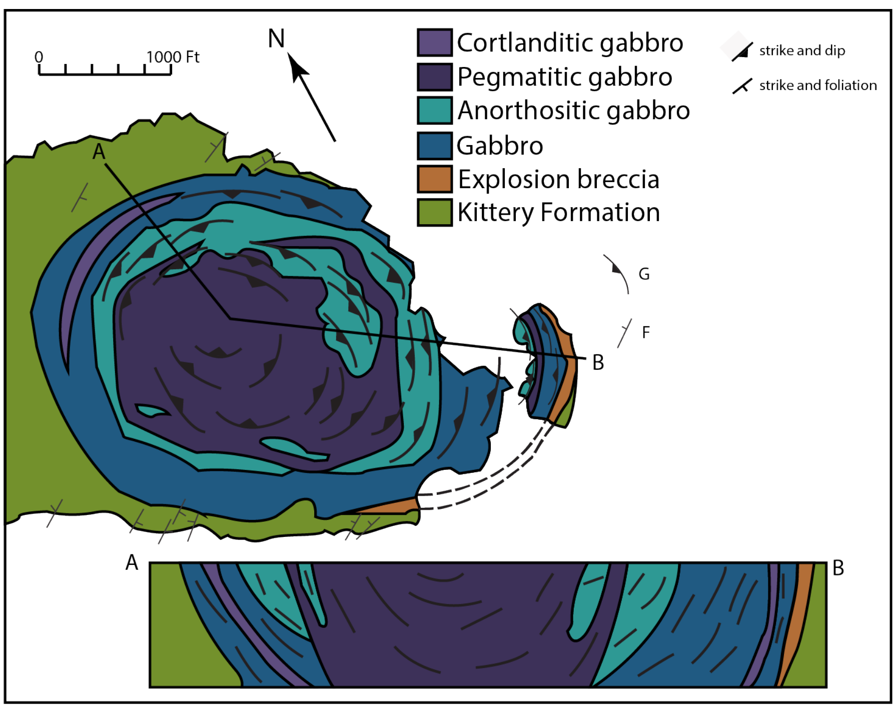


Figure 41-3. Geologic map of the Cape Neddick Gabbro, York, ME (After Hussey, 2015). Cortlanditic gabbro is an olivine, pyroxene, amphibole bearing gabbro, anorthositic gabbro is a feldspar-rich gabbro, and explosion breccia contains fragments of broken rock owing the sudden explosive emplacement of magma. Strike and dip and strike and foliation symbols show the 3 dimensional orientation of structures in the gabbro.

Walking Directions

All these photos were taken very near the parking lot. GPS coordinates will guide you to these locations.

Figure 41-4. Typical gabbro of the Cape Neddick complex. The dark mineral is pyroxene, the lighter mineral is plagioclase feldspar.

Figure 41-5. Faint layering is present in many locations in the Cape Neddick gabbro (N 43º 09.958’ W 070º 35.592’).

On the Outcrop

At various locations in Sohier Park, one can observe a faint layering in the pluton (Figure 41-5). The thin, lighter layers are mainly plagioclase, the darker layers are a mix of both plagioclase and pyroxene.

At advanced stages of crystallization of the magma, coarser-grained pegmatites formed. These are most easily spotted by the abundance of amphibole crystals in the dikes. The pegmatitic magma was richer in water than the gabbro, hence amphibole, a mineral that contains about 3 weight percent water, forms instead of pyroxene. Some of the pegmatites are quite thin as seen at N 43º 09.951' W 070º 35.569' (Figure 41-6).

Also present in the pluton are late-stage, aplitic dikes (Figure 41-7). Aplites are fine grained, very evolved rocks that represent the last gasp of liquid in plutons. These are finer-grained, amphibole-poor, felsic dikes consisting predominantly of feldspars and quartz. As the gabbro crystallized, it solidified to large amounts of crystals with minor, interstitial melt among the crystals. That interstitial melt is sometimes squeezed out from the crystals to form these very evolved, light colored aplites.

As seen at Stops (21a, 35c, and 39), basaltic dikes also cut the Cape Neddick Gabbro (Figure 41-8). However, because the Gabbro is Cretaceous, about 116 million years old, it is much younger than any of the dikes seen previously that are associated with the breakup of Pangea and the formation of the Atlantic Ocean basin. This dike clearly cuts the gabbro, containing gabbroic xenoliths (foreign rock), hence it is younger than the pluton (Figure 41-8).

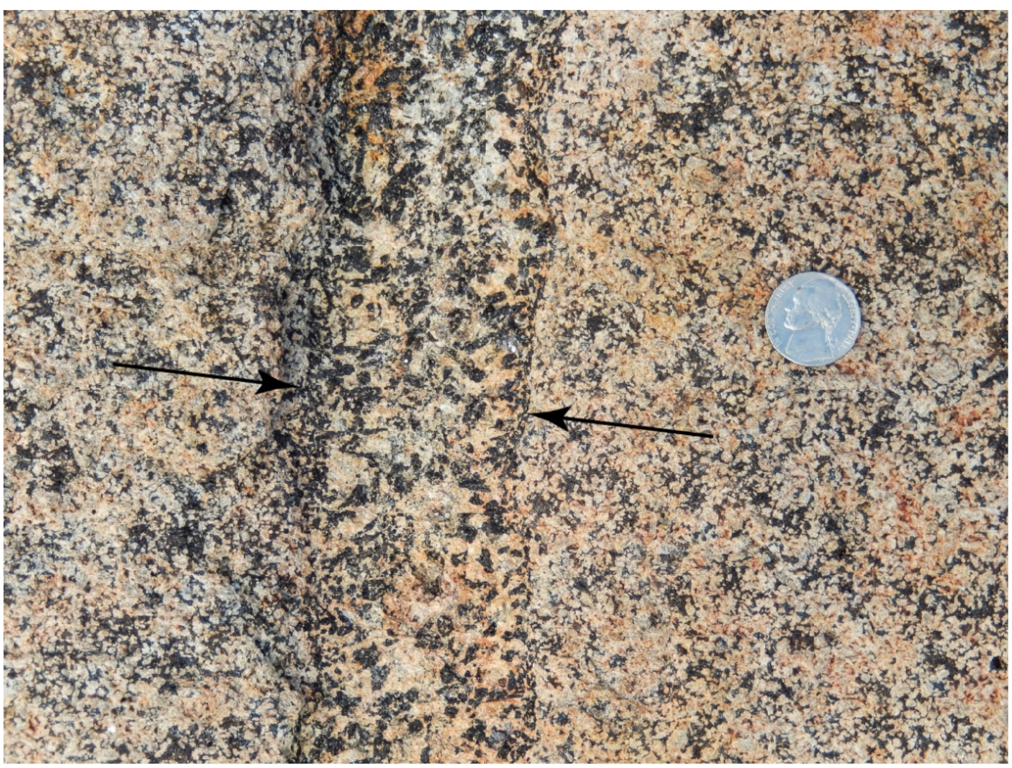


Figure 41-6. Numerous pegmatitic dikes also cut the gabbro. In this case, the pegmatite (shown between the arrows) is amphibole-rich (N 43º 09.957' W 070º 35.595').

Figure 41-7. Late-stage, aplitic dikes cut the gabbro (N 43º 09.961’ W 070º 35.609’).

Figure 41-8. Gabbroic xenoliths in late-stage dike.(N 43º 09.949’ W 070º 35.567’).

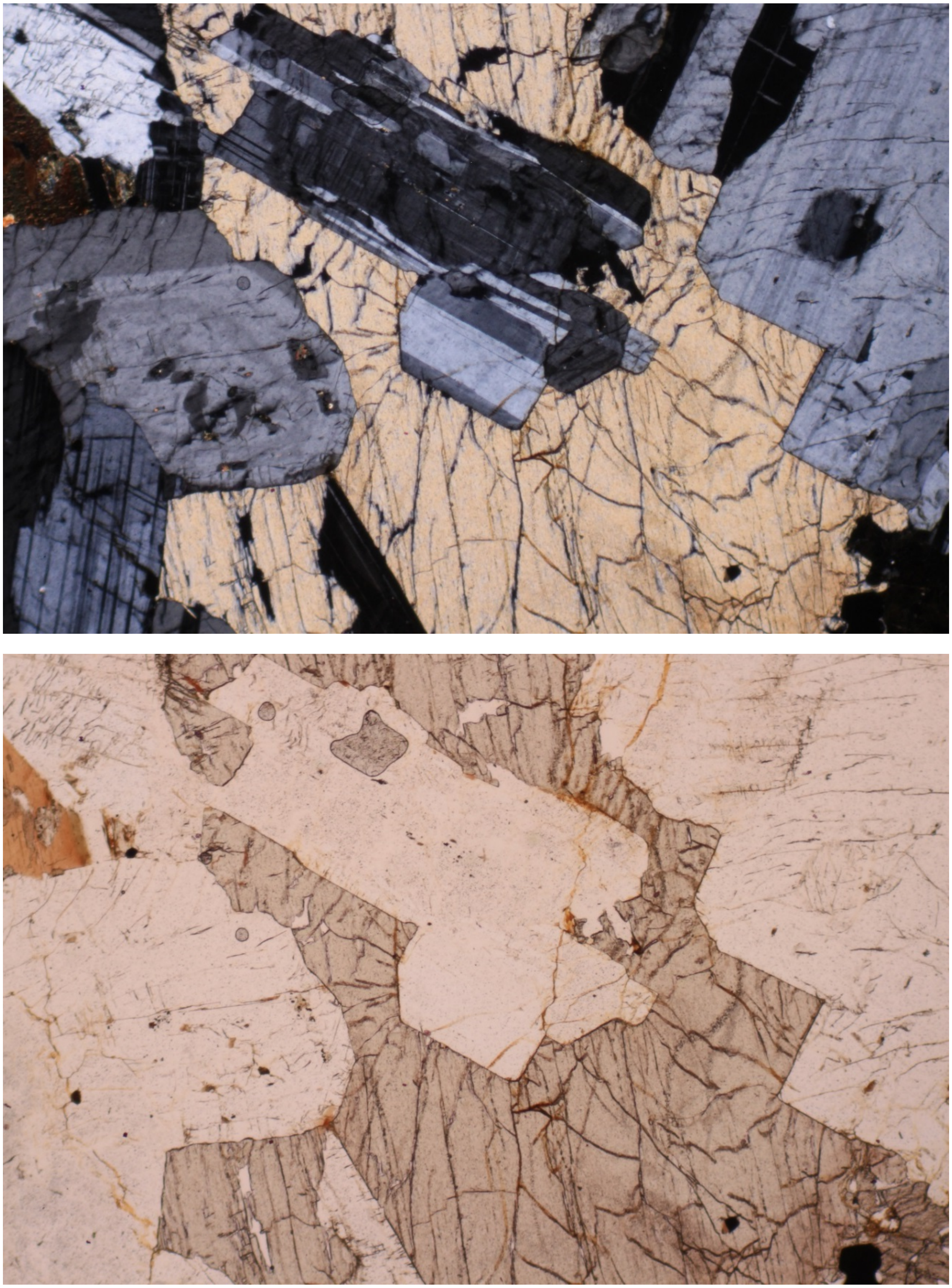


Figure 41-9. Crosses polarized light (upper photo) and plane polarized light (lower photo) photomicrographs of the Cape Neddick Gabbro. The gray to black mineral of the upper photo is plagioclase feldspar, the pale brown is pyroxene. In the middle of the photos, a plagioclase crystal is included in a larger, light brown pyroxene grain.

42. Christmas Cove Dike, Central Atlantic Magmatic Province, South Bristol, ME

The 201 Ma Christmas Cove Dike is a segment of a Central Atlantic Magmatic Province dike that extends from Connecticut to Maritime Canada. The Central Atlantic Magmatic Province resulted from the outpouring of thousands of cubic miles of basaltic magma during the rifting of Rodinea that initiated the formation of the Atlantic Ocean basin (Figure O-15). Massive extinctions occurred at this time, marking the end of the Triassic Period and the beginning of the Jurassic. The dike is called the Higganum Dike in Connecticut, the Holden Dike in Massachusetts, the Onway Dike in New Hampshire, and the Christmas Cove dike at this location. At the time that various geologists mapped the dike in the different states, the age was not clearly defined and the mappers were unaware that their dike was a continuation of a single dike that extends the length of New England, hence they gave regional names to the dike segments they mapped. Since then, the similar ages and identical chemistry of each dike segment made it apparent that the dike is actually continuous for over 900 miles. This dike fed lava flows that erupted and covered such vast area that such eruptions are called flood basalts. This is the same dike as seen at Stop 22 at Onway, NH.

Driving Directions

Take RT 1 to RT 1B to Damariscotta, ME. From Damariscotta, follow RT 129 to South Bristol. About 1.2 miles for the center of South Bristol, there is a narrow stretch of land with water on both sides of the road, giving a "causeway" appearance to the road (Figure 42-1). Outcrops are on the left, to the northwest side of the "causeway". Park on the east side of the road.

Figure 42-1. Map of Christmas Cove showing the location of outcrops for this stop.

Figure 42-2 Outcrop of the Christmas Cove dike (N 43º 50.800’ W 069º 33.073’).

On the Outcrop

It’s hard to determine the thickness of the dike at Christmas Cove because both contacts are not exposed here (Figure 42-2). But the dike commonly exhibits columnar jointing (columns formed by contraction during cooling of the magma) that are perpendicular to the dike walls, giving information on the orientation of the dike. Elsewhere in South Bristol, the dike is about 75 feet thick, similar to the NH portion of the dike at Stop 22. Regardless, a massive volume of magma moved through this dike and similar CAMP dikes, feeding one of the largest flood basalt provinces in the geologic record.

The drilled hole shown in Figure 42-3 marks the sample location for a paleomagnetic study. Earth's magnetic field extends from the Earth's interior, out the South Magnetic Pole and down at the North Magnetic Pole. The magnetic field is generated by electric currents produced by the motion of convection currents of molten iron in the Earth's outer core. The angle of inclination of the magnetic field varies with latitude. It points straight down (90º) at the North Magnetic Pole in the northern hemisphere and declines to progressively lower angles as the latitude decreases until it is horizontal (0°) at the equator. It continues to rotate until it is straight up at the South Magnetic Pole.

Figure 42-3. Close-up photo of the Christmas Cove dike. The fine-grained nature of the dike may not yield much to look at in the field, but the photomicrographs presented of the Onway Dike of Stop 26 show the rock to contain abundant pyroxene and feldspar crystals.

Geophysicists can determine the latitude of the eruption of an igneous rock by determining the paleomagnetic signature. Iron-titanium oxide minerals in igneous rocks may preserve the inclination of the Earth's magnetic field when the rocks cool through about 580° C. Igneous rocks are completely solidified at this temperature, hence, the mineral grains are not rotated physically to align with the Earth's field, but rather they record the orientation of that field. These measurements require a horizontal orientation of the igneous lava flow; orogenic events deform and tilt rocks from their original orientations. Fortunately, sediments are interbedded with CAMP lava flows in the Hartford Basin of Connecticut, enabling original horizontality to be determined. These studies reveal that 201 million years ago when the supercontinent Pangea began to rift and the flood basalts of the Central Atlantic Magmatic Province erupted, this portion of New England was located about 25^{O} North latitude.

In the case of the Christmas Cove Dike and its extension, the Onway Dike of Stop 22, the drill holes seen at the outcrops were not used to infer paleolatitude because one cannot be certain of the emplacement orientation of the dike. Subsequent tilting could have rotated the dike from its original orientation. The paleomagnetic properties of the dikes were used to determine flow direction of the magma: did the magma ascent vertically, arising from a very large scale melt zone or did if flow laterally, with the dike radiating from a single location? It was found that the dike has a strong lateral flow component, having flowed from the southwest to the Christmas Cove location and farther into Maritime Canada.

GLOSSARY OF GEOLOGIC TERMS

Acadian Orogeny: A mountain building event resulting from the collision of the microcontinent Avalonia with Laurentia. The orogeny lasted between 410 to 390 Ma.
Alleghanian Orogeny: A mountain building event resulting from the collision of the Gondwanan continent with Laurentia to form the supercontinent Pangea. The orogeny lasted between 310 to 280 Ma.
alluvial fan: A deposit of sediments that originate from a point source in a mountainous terrane at the apex of the fan and spread laterally to form a fan shaped deposit of coarse-grained sediment onto a flatter, adjacent plain.
ammonite: An extinct life form of the class Cephalopoda that is related to modern octopuses, squid, and cuttlefish. The animal inhabited the most outward chamber of a spiral shaped shell.
amphibolite: A metamorphic rock consisting dominantly of amphibole (usually hornblende) and plagioclase feldspar. The protolith (original rock prior to metamorphism) is commonly basalt.
andalusite: The low pressure form of Al_2SiO_5. Andalusite typically forms as elongated, prismatic crystals and is most common in medium temperature metamorphic schists and gneisses. See Figure I-6 for its pressure-temperature stability field.
anticlinorium: A large, regional scale anticline upon which minor folds are superimposed. An anticline is an arch-like fold with the oldest layers exposed in the center and younger layers on the limbs.
anorogenic: A geologic event, usually related to magmatism, that is not related to a mountain-building event.
asthenosphere: The mechanically weak, ductility deformable layer of the Earth's upper mantle. It is below the lithosphere at depths of ~ 50-120 miles below the Earth's surface.
augen gneiss: A gneiss with deformed feldspar crystals that are tapered on each end, forming eye-shaped porphyroblasts. Augen is German word for eyes.
Avalonia: A microcontinental fragment rifted off Gondwana during the Silurian period. Collision of Avalonia with Laurentia caused the Late Devonian, Acadian Orogeny. Also known as Avalon.
basalt: A volcanic igneous rock that may consist of quenched glass with small, crystals of plagioclase feldspar, pyroxene, and olivine. Phenocrysts (larger crystals) of these same minerals may be present. Basalt is chemically the same to its intrusive equivalent, gabbro.
basement: Crystalline rocks that underlie the oldest sedimentary rocks in an area. Basement complexes typically consist of igneous and/or high grade metamorphic rocks that have younger, less metamorphosed and/or sedimentary cover rocks above them.
beryl: A beryllium aluminium silicate Be_3Al_2(Si6O18). Most commonly found in grantic pegmatites.
biotite: A black or dark brown mica that is rich in iron and/or magnesium. It is common in granitic rocks and especially in schists.
blueschist: A metamorphic rock subjected to high pressures at relatively low temperatures. The rock is characterized by the blue-colored amphibole glaucophane.
brachiopod: Marine animals with hard shells, somewhat similar to clams, but the two shells are different sizes.
breccia: A coarse-grained rock with angular fragments set in a finer-grained matrix.

bryozoan: A phylum of aquatic invertebrate animals that live in colonies of interconnected individuals. They are sometimes called moss animals because the colonies sometimes resemble mosses.
boudin: An elongated, sausage-shaped rock resulting from stretching of a more rigid rock within are more ductile host rock.
calcite: A calcium carbonate, $CaCO_3$, that forms as a chemical precipitate, commonly in sea water. It is the dominant component in most limestones and marbles.
carbonate: A group of minerals with Ca, Mg, or Fe bonded to CO_3 ions. Calcite is the $CaCO_3$ version, dolomite is $CaMg(CO_3)_2$, the common constituent of dolostone.
cephalopods: Members of the molluscan class Cephalopoda such as squids, octopuses, nautiluses, and ammonites. They are marine animals that may inhabit a shell (e.g., nautiluses, ammonites), but all have arms and tentacles.
columnar jointing: Prismatic, hexagonal or pentagonal columns that form by contraction during cooling. Most commonly form in basaltic lava flows, but are present in other lavas as well.
complex: An assemblage of rocks that are structurally related but may have different ages or origins.
conglomerate: A coarse-grained, clastic sedimentary rock with clasts greater than 2 mm in diameter.
continental arc: An arc composed of a chain of volcanoes or their plutonic equivalents that formed by subduction of an oceanic plate under the edge of a continent. The Andes of South America and the Cascade Mountains of northern Californian to Washington are present-day continental arcs. The Sierra Nevada Batholith represents the plutonic component of a continental arc.
crenulation cleavage: A fabric in metamorphic rocks that were subjected to two or more stress directions causing an overprinting of the earlier planer fabric by the younger fabric.
Cretaceous: A period of the Mesozoic era lasting from 145 to 66 Ma.
Devonian: A period of the Paleozoic era lasting from 419.2 to 358.9 Ma.
diamictite: A conglomerate with clasts that are supported by a finer-grained matrix such that the clasts generally are not in contact with each other.
dike: A sheet-like intrusion of igneous rock that cuts across the bedding or foliation of the rocks they intrude.
diorite: A plutonic rock with only minor quartz and potassium feldspar, but with abundant plagioclase feldspar. Typically, the remainder of the rock mainly consists of amphibole, pyroxene, or biotite. See Figure I-5 for plutonic rock classification scheme.
dolostone: A carbonate rock dominated by dolomite, the $CaMg(CO_3)_2$ carbonate mineral.
dome: A circular to elliptical shaped uplift with the flanking rocks angling away from the central, structural high**.**
dorsal zone: A zone where rocks have been pushed up from a central zone and thrust in opposite directions away from that central zone. Also known as a flower structure.
drift stage: Subsequent to continental rifting, the continent drifts away from the spreading center to form a passive margin.
dunite: An ultramafic rock dominated by 90% or greater amounts of the mineral olivine $(Mg,Fe)SiO_4$. Most dunites are derived from the mantle.
enclave: A piece of rock that is surrounded by an igneous host rock of a different composition., e.g., a diorite in granite.

epidote: A hydrous, iron, calcium silicate{$Ca_2Al_2(Fe^{3+};Al)(SiO_4)(Si_2O_7)O(OH)$} that is characterized by a pistachio green color. It typically forms resulting from the alteration of other minerals by hydrothermal fluids.
facies: Rocks of similar composition, e.g., basalts, that were subjected to the same pressure-temperature conditions, will have the same metamorphic minerals. Different assemblages of minerals will form at different pressure-temperature (P-T) conditions. These different assemblages have been used to define specific P-T zones or facies in P-T space. See Figure I-6.
fiamme: Flame-like structures in pyroclastic volcanic rocks. They are the result of either compressed pumice fragments or result from small inclusions of magma of different composition than the host.
flood basalt: Massive eruptions of basaltic lavas that erupt during continental rifting events.
flute cast: Sedimentary structures on the botton of certain sedimentary beds. They result from scouring of a sediment-laden current over an unconsolidated sedimentary bed. Also known as sole markings.
footwall: The underlying side of a fault.
gabbro: An igneous rock that is the plutonic equivalent to basalt. It consists primarily of plagioclase feldspar, pyroxene and possibly olivine.
Ganderia: A microcontinental fragment rifted off Gondwana during the Ordovician period. Collision of Ganderia with Laurentia caused the Salinic Orogeny. The Massabesic Gneiss Complex of New Hampshire and many coastal Maine rocks are Ganderian. Also known as Gander.
garnet: A silicate mineral with a wide range of compositions, but in the metamorphic rocks of northern New England, the almandine ($Fe_3Al_2Si_3O_{12}$) version of garnet is dominant. It is reddish to reddish brown in color and is especially common in schists.
gastropods: A large group of mollusks characterized by a single, often coiled, shell. Modern representatives include snails.
glaucophane: A bluish colored amphibole that is produced under high pressure conditions. Its presence gives the name blueschist to high pressure metamorphic rocks and the blueschist facies.
gneiss: A banded metamorphic rock with lighter colored layers that are typically rich in quartz and feldspar alternating with darker bands that may be dominated by micas and/or amphibole. Gneisses with igneous protoliths are orthogneissses. Paragneisses have sedimentary protoliths.
Gondwana: A supercontinent that rifted to form the present-day continents of Africa, South America, Australia, and Antarctica. India is also a fragment of Gondwana. The collision of Gondwana with Laurentia formed the supercontinent Pangea.
graben: A fault-bounded valley that forms during rifting where the valley floor drops with respect to the adjacent crust.
granite: A plutonic igneous rock consisting of approximately equal proportions of potassium feldspar, plagioclase feldspar and quartz. The rock may also contain smaller amounts of mica and/or amphibole. See Figure I-5 for plutonic rock classification scheme.
granodiorite: A plutonic igneous rock consisting of quartz and greater proportions of plagioclase feldspar than potassium feldspar. The rock may also contain smaller amounts of mica and/or amphibole. See Figure I-5 for plutonic rock classification scheme.
granophyre: A rock texture consisting of microscopic intergrowths of quartz and feldspar.
granulite: A metamorphic rock formed under high temperatures and potentially, over a range of pressure conditions. See granulite facies in Figure I-6.

graphic granite: A coarse-grained granite with large crystals of quartz and feldspar that are intergrown to resemble cuneiform writing.
greenschist: Metamorphic rocks formed at relatively low temperatures and pressures, generally between 230-500ºC and 0.2-.08 GPa (See Figure I-6).
Grenville Orogeny: A late Precambrian orogenic event that formed along the eastern portions of Laurentia (present day coordinates) between 1.2 and 1.0 billion years ago. Grenville rocks are exposed in the Green Mountains of Vermont.
greywacke: A texturally immature sedimentary rock consisting of angular grains of quartz, feldspar and lithic fragments.
hanging wall: The overlying side of a fault.
harzburgite: An ultramafic igneous rock consisting of mainly olivine and orthopyroxene.
hornblende: A member of the amphibole family with the general formula of $(Ca,Na)_{2-3}(Mg,Fe,Al)_5(Al,Si)_8O_{22}(OH,F)_2$.
hornfels: A metamorphic rock formed by the contact of an igneous intrusion. Hornfels are noted for their hardness compared to the original protolith.
hydrothermal: Hot water, usually related to a magmatic origin.
Iapetus Ocean: Rifting of the supercontinent Rodinia about 550 Ma produced an ocean basin occupied by the Iapetus Ocean located between Laurentia and the microcontinent of Avalonia. Subsequent subduction consumed the basin and led to the demise of the Iapetus Ocean. The name was derived from the Titan Iapetus, who, according to Greek mythology, was the father of Atlas after whom the Atlantic Ocean was named.
Jurassic: A period of the Mesozoic era ranging from 201.3 to 145 Ma.
hematite: A mineral with the composition of Fe_2O_3.
immature sediments: Sediments with poorly sorted and angular grains, some of which are unstable at the Earth's surface.
island arc: A volcanic arc composed of a chain of volcanoes formed by subduction of an oceanic plate under the edge of another oceanic plate. The Philippine Islands and the Japanese Archipelago are present-day island arcs**.**
kyanite: : The high pressure form of Al_2SiO_5. Kyanite typically forms bluish blades. See Figure I-6 for its pressure-temperature stability field.
Laurentia: A name for the Proterozoic to Paleozoic ancestor of North America.
leucosome: The light colored portion of a migmatite that represents the partial melt fraction of the rock.
lithosphere: The outer rigid layer of the Earth, includes the crust and the uppermost mantle**.**
lherzolite: An ultramafic igneous rock consisting mainly of olivine, clinopyroxene and orthopyroxene.
mafic: A term pertaining to igneous rocks that are rich in ferromagnesian elements, e.g., basalts and gabbros.
miarolitic cavity: A small irregular-shaped cavity in plutonic rocks, representing a gas bubble, commonly with crystals that grew into the open space.
mantle: A zone of the Earth, extending between the crust and the core at a depth of ~ 2100 miles. It is divided into an upper and lower mantle.
marble: A metamorphic rock consisting mainly of calcium carbonate minerals.
mature sediments: Sediments consisting mainly of well-rounded particles and minerals that are stable at the Earth's surface.

Meguma: A microcontinental fragment of Gonwana that separated during the Paleozoic era and was accreted to Laurentia during the Neoacadian Orogeny. The terrane is not exposed in New England but is present in Nova Scotia.
mélange: A heterogeneous body consisting of a mixture of rock materials.
melanosome: The dark portion of a migmatite, representing portions of the rock after melt extraction.
mid-oceanic ridge: An underwater mountain range formed where two tectonic plates diverge. They are sites of young magmatism where new oceanic crust is created.
migmatite: A rock that has undergone partial melting, consisting of a magmatic component (leucosome) and a darker, residual portion resulting from melt extraction (melanosome). Some migmatites contain paleosomes, the initial rock that did not partially melt.
mullions: A wave-like structure formed along a fault plane by movement of one body of rock past the other.
muscovite: A colorless to silver colored mica that is abundant in many metamorphic schists and in some granites such as the Concord Granite of New Hampshire.
nappe: A large sheet of rock that has moved horizontally over younger rocks.
Neoacadian Orogeny: A mountain building event resulting from the collision of the microcontinent Meguma with Laurentia. The orogeny lasted between ~370-350 Ma.
nonconformity: An erosional surface between sedimentary rocks and underlying plutonic or massive metamorphic rocks.
normal fault: A fault where the hanging wall dropped down with respect to the footwall block.
oligomict conglomerate: A coarse-grained sedimentary rock where the clasts consist of a single rock type.
olivine: An iron, magnesium silicate ($(Mg,Fe)_2SiO_4$) that commonly crystallizes in basaltic magmas. It is easily altered to serpentine under hydrous, low grade metamorphic conditions.
ophiolite: A section of the Earth's oceanic crust and the adjacent upper mantle that was thrust onto continental crust at continental margins. It consists, from top to bottom, of oceanic sediments, pillow basalts, sheeted dikes, gabbros, and peridotites.
orogeny: A term indicating the process of mountain formation.
orthogneiss: A felsic gneiss whose original rock, prior to metamorphism, was igneous.
paleocurrent indicator: Sedimentary structures that give indications of current directions when the sediments were being deposited.
paleomagnetism: The study of the magnetic properties of rocks to determine the strength and direction of the Earth's magnetic field in the past.
paleosome: The unmelted portion of a migmatite.
Pangea: A supercontinent that formed during the late Paleozoic about 335 million years ago. It began to rift about 200 million years ago.
paragneiss: A high grade, metamorphic rock derived from sedimentary protoliths.
passive margin: A tectonically inactive continental margin that marks the transition between the continental and oceanic lithosphere.
pegmatite: A very coarse-grained igneous rock, usually of granitic composition, consisting mainly of feldspars and quartz. May also contain other rare, economically valuable minerals such as tourmaline.
peridotite: An ultramafic igneous rock consisting of mainly olivine and pyroxenes.
Permian: A period of the Paleozoic era spanning the time between 298.9 to 251.9 Ma.

perthite: A potassium-rich alkali feldspar ($KAlSi_3O_8$) that cooled slowly and exsolved lamellae of sodic alkali feldspar ($NaAlSi_3O_8$). The thickness of the lamellae varies, and is sometimes visible to the naked eye.
Phanerozoic eon: The time span comprising the Paleozoic, Mesozoic, and Cenozoic eras.
phenocryst: A crystal in an igneous rock that is usually larger than the groundmass of the rock. Typically larger than 5 mm in diameter.
phyllite: A fine-grained, foliated metamorphic rock. It contains fine-grained micas that are aligned in a preferred orientation, giving a sheen to the rock.
pillow basalt: Lavas with pillow-shaped structures that form as the basalt comes in contact with water.
plagioclase: A type of feldspar that ranges in composition between albite ($NaAlSi_3O_8$) and anorthite ($CaAl_2Si2O_8$).
pluton: A body of igneous rock that intruded and crystallized at depth.
polymict conglomerate: A type of conglomerate with clasts from a variety of different rock types.
porphyroblast: A relatively large mineral formed during metamorphism.
potassium feldspar: The potassium member of the feldspar group ($KAlSi_3O_8$). It is a major constituent of granite and is the dominant mineral in syenites.
Proterozioc: An eon of the Precambrian that lasted from 2500 to 541 Ma.
protolith: The original rock before it was metamorphosed, e.g., clay-rich sedimentary rocks are common protoliths to metamorphic schists.
pseudotachylite: A rock representing melted rock formed by friction as rocks slide past each other along faults during earthquake producing events. Its glassy nature renders it darker than the unmelted rocks adjacent to it.
pumice: A light colored, vesicle-rich rock that represents a bubble-rich froth at the top of magma chambers.
pyroclastic eruption: A rapidly moving current of hot gas and volcanic ash that descends the slopes of a volcano at speeds of 60 mph. Also known as nuée ardente, French for "burning cloud".
pyroxene: A magnesium, iron and in some pyroxenes, calcium-rich silicate. It tends to form as an igneous mineral at higher temperatures than amphibole. With increasing water concentration in an evolving magma, amphibole tends to replace pyroxene.
quartzite: A metamorphic rock consisting of >90% quartz.
quartz syenite: A plutonic igneous rock that is dominated by potassium feldspar, less than 35% plagioclase, and between 5 and 20% quartz. See Figure I-5 for plutonic rock classification scheme.
rapakivi granite: A granite containing rounded crystals of orthoclase ($KAlSi_3O_8$) that are mantled by plagioclase feldspar ($NaAlSi_3O_8$ to $CaAl_2Si2O_8$).
reverse fault: A fault where the hanging wall moves over the footwall, results from compression.
Rheic Ocean: An ocean outboard of Avalonia, Ganderia, and Meguma. Following the consumption of the Iapetus Ocean, the Rheic Ocean was the main ocean between the modified Laurentian continent and Gondwana. The Rheic Ocean was consumed by subduction that closed the ocean basin during the Alleghanian Orogeny. It was named after Rhea, the sister of Iapetus who was the father of Atlas of Greek mythology.
Rodinia: A Proterozoic supercontinent formed during the Grenville Orogeny.
rhyolite: The volcanic equivalent to granite.
Salinic Orogeny: A Paleozoic mountain building event caused by the collision of Ganderia with Laurentia.

schist: A metamorphic rock with a platy fabric defined by the common alignment of abundant micas.
serpentine: A hydrous metamorphic mineral commonly formed by the addition of water under low grade metamorphic conditions to olivine-rich protoliths.
serpentinite: A metamorphic rock consisting of greater than 90% serpentine.
sheeted dike complex: A body of rock formed when dikes continually intrude extending crust and sequentially younger dikes intrude earlier emplaced dikes.
shield: Large, low-lying, stable regions of the Earth's crust composed of Precambrian crystalline rocks.
sillimanite: The high temperature form of Al_2SiO_5. Sillimanite typically forms elongated, needle-like crystals that are usually microscopic. It is most common in medium temperature metamorphic schists and gneisses. See Figure I-6 for its pressure-temperature stability field.
Silurian: A geologic period of the Paleozoic era, lasting between 443.8 and 419.2 Ma.
sole mark: A sedimentary structure found at the base of some sedimentary layers. They are formed as turbidity currents flow over and scour grooves in the underlying sedimentary sheet. As the current comes to rest, sediments from the turbidity current fill in the scour marks, forming elongated structures at the base of the overlying layer.
staurolite: An iron, aluminum silicate that forms in aluminum-rich schists in the amphibolite facies.
strike-slip fault: A fault where to two blocks slide horizontally past each other.
stromatoporoids: A dominant type of reef-building, calcareous sponge that lived from the Ordovician through the Devonian. A Mesozoic form was a major reef-builder during the Cretaceous.
stoping: A method of magma intrusion where blocks of the surrounding rock are broken and sink down into the pluton, allowing the magma to move upward.
subduction: The descent of one tectonic plate below another.
successor basin: Sedimentary basins formed after mountain formation as the crust weakens and extends.
supercontinent: A large continent formed by the amalgamation of several continents.
syenite: A plutonic igneous rock that is dominated by potassium feldspar, less than 35% plagioclase, and less than 5% quartz. See Figure I-5 for plutonic rock classification scheme.
tabulate corals: An extinct form of coral consisting of hexagonal cells, giving a honeycomb appearance. They lived during the Paleozoic.
Taconic Orogeny: A mountain building event caused by the collision of an island arc with the Laurentian margin. The orogeny lasted between 470-450 Ma.
terrane: A fragment of crust broken from one tectonic plate and accreted to another during an orogenic event at convergent margins.
thrust fault: A type of reverse fault where the fault angle is less than 45 degrees. The hanging wall is pushed over the footwall, often for several miles.
tonalite: A plutonic rock with minor potassium feldspar, abundant plagioclase feldspar, but more quartz than diorite. Typically, the remainder of the rock mainly consists of amphibole and/or biotite. See Figure I-5 for plutonic rock classification scheme.
tourmaline: An igneous and metamorphic mineral that exhibits a wide range of compositions, but in the schists of northern New England, the iron-rich variety schorl ($NaFe^{2+}{}_3Al_6(BO_3)_3Si_6O_{18}(OH)_4$) is most common. Gem quality elbaite ($Na(Li_{1.5},Al_{1.5})Al_6Si_6O_{18}(BO_3)_3(OH)_4$) occurs in pegmatites in New Hampshire and Maine.

Triassic: The first period of the Mesozoic era between 251.9 and 201.3 million years ago.
trough: An elongated sedimentary basin.
turbidite: A sedimentary deposit of a turbidity current, or sediment-laden, marine avalanche, characterized by erosional contacts with the underling bed and upward grading of grain sized from coarse to fine.
ultramafic: Igneous rocks that are rich in Mg and Fe, generally consisting of olivine and pyroxenes. Ultramafic rocks include peridotites, pyroxenites, and dunites.
unconformity: A gap in the stratigraphic record between two rock masses or strata of different ages. The gap represents a time of erosion between the two rock masses.
xenolith: A fragment of foreign rock in igneous host rock. Typically, these are fragments of the surrounding country rock that broke off and were incorporated as rafts in the granitic magma.
zircon: A zirconium silicate ($ZrSiO_4$) that is a common assessor mineral in igneous, metamorphic, and sedimentary rocks. This mineral is especially useful because it contains thorium and uranium that decay to lead, the measurement of which allows the age of formation of the crystal.

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