Bedrock geology of the Presidential Range, New Hampshire

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INTRODUCTION

This field guide outlines the bedrock geology in the alpine zone of the Presidential Range, New Hampshire and describes some of the significant and readily accessible outcrops in the vicinity of Mt. Washington. The details of the five year project to map the geology in the alpine zone has been presented in the Geological Society of America Bulletin by Eusden et al. (1996). For a complete explanation of the alpine zone geology, please refer to this paper which is accompanied by oversized figures showing the geologic map, cross sections, and maps of several structural fabrics. This field guide differs from Eusden et al. (1996) in that some of the more interesting local and historical details of the geology are expanded upon here.

The Presidential Range, with its unique natural environment, has always been a haven for scientific exploration. Many topographical features have been named after scientific researchers. Jackson, Huntington, Hitchcock, and Agassiz were among the researchers who worked in the Presidential Range during the 1800's. Many geologists have since studied various aspects of the geology and research still continues today. More recent studies done this century, including the classic works of Billings and Fowler-Billings, are discussed below.

Previous Work

Stratigraphy. Billings (1941) assigned the metamorphic rocks of the Mt. Washington area formational names based on correlations made to the Littleton-Mount Moosilauke area of New Hampshire. The youngest unit was the Devonian Littleton Formation, a quartzite and schist. Beneath the Littleton was the Silurian Fitch Formation, which was described as a thin, discontinuous unit derived from impure dolomites. The upper Ordovician Partridge Formation was the oldest metamorphic unit with local schists and quartzites, but largely a gneiss (Billings, 1941). Billings et al. (1946) reassigned all the metamorphic rocks to the Devonian Littleton Formation. They introduced the Boot member as a replacement for the Fitch. The rest of the Littleton was designated into an upper (the former Littleton schists and quartzites) and lower unit (the former Partridge gneisses). Billings and Fowler-Billings (1975) and Billings et al. (1979) again discussed the possibility that the Boot member and strata below it could be Silurian in age, but kept the rocks assigned to the Devonian Littleton Formation.

Hatch et al. (1983) correlated the metasedimentary rocks in the Pinkham Notch area to a stratigraphy described by Moench (1971) in the Rangeley, Maine area, along strike, 40 miles northeast. Hatch et al. (1983) extended the use of the Silurian formational names, Rangeley, Perry Mountain, Smalls Falls, and Madrid, southwest into New Hampshire. The Devonian Littleton Formation was retained in New Hampshire and is correlative to the Devonian Carrabassett, Hildreth, and Seboomook Formations of Maine (Hatch and Moench, 1984).

In the Presidential Range, the Silurian Perry Mountain Formation is missing from the maps of Hatch and Moench (1984), Hatch and Wall (1986), Wall (1988), and Lyons et al. (1992) due to either non-deposition, erosion, or pre-metamorphic gravity slides. Allen (1992) does include a small section of Perry Mountain in his stratigraphic section of the Carter Dome-Wild River region and the Pinkham Notch area adjacent to the Presidential Range. These rocks are assigned to the Perry Mountain due to the abundance of quartzites and position between the Rangeley and the Smalls Falls Formations.

The Silurian Rangeley, Perry Mountain, and Smalls Falls Formations are believed to have a southeast sediment transport direction, shed from the eroding Bronson Hill arc (Hanson and Bradley, 1989). The sediments of
the Madrid Formation were transported along the axis of the basin, towards the southwest (Hanson and Bradley, 1989). The Devonian Littleton sediments had an overall northwesterly transport direction, toward the margin of the pre-Acadian North American continent, shed from the west (present coordinates) margin of Avalon (Hanson and Bradley, 1993).

The basin discussed above essentially describes an extensional, passive tectonic setting throughout the deposition of the Siluro-Devonian sediments. Although this model has been widely accepted, it is often difficult to rationalize tectonically, as the Acadian orogeny occurred immediately after the deposition of the Littleton. In fact, recent studies by Bradley and Hanson (1989) and Hanson and Bradley (1989), have suggested that the upper Madrid and the Carrabassett Formations in central Maine were deposited in an active trench system causing the formations to break into mélanges and/or olistostromes. In the southern Presidential Range, calc-silicate rip-up clasts in the Rangeley Formation have been similarly described by Guthrie and Burnham (1985) as evidence of a debris flow in a sedimentary basin, possibly of olistostromal origin. Whether these are the tectonic mélanges or sedimentary olistostromes is debatable, and, whether these sedimentary facies are interpreted as a result of extension or convergence is critical to plate tectonic models for the Acadian.

Structural Geology. Billings (1941) and Billings et al. (1979) described the structure of the Presidential Range as being the product of a single phase of both major and minor asymmetrical, plunging, en echelon folding. The minor folds are abundant throughout the range and have amplitudes and wavelengths that range from a few inches to many feet. Axial planes usually dip steeply to the northwest. In Billings’ (1941) model, Mt. Washington culminates in a major asymmetric anticline, without any large-scale overturning or nappe-like recumbent structures.

Hatch and Moench (1984) recognized multiple phases of deformation in the metasedimentary rocks exposed in the White Mountain National Forest. A pervasive schistosity, in many cases axial planar to early isoclines, is characteristically refolded by more open, late folds with north striking, steeply dipping, axial plane cleavage surfaces (Hatch and Moench, 1984). A similar account of the structures is given by Hatch and Wall (1986) for the Pinkham Notch area. Hatch and Wall (1986) have also documented the existence of early isoclines by repeated reversals of topping indicators (graded beds). Neither facing directions nor vergences for these two phases of deformation are reported by Hatch and Moench (1984) or Hatch and Wall (1986).

In a regional compilation for the state of New Hampshire’s bedrock geologic map, Lyons et al. (1992) further modified the maps of Hatch and Moench (1984) and Billings (1941). The contact between the Smalls Falls/Madrid and Rangeley formations is shown as a normal fault with the Littleton/Smalls Falls/Madrid stratigraphy in the upper plate. The Mount Clay area is shown as an isolated klippe of Rangeley formation. Presumably, the Silurian formations, that were once connected by Hatch and Moench (1984) and Billings (1941), are now separated by Lyons et al. (1992) because of a reassessment of the bedrock geology on the west flank of the Presidential Range.

Allen (1992) working in the Pinkham Notch area suggests that the original structure of the meta-sedimentary rocks now migmatized appears to have been one of large scale, east vergent isoclinal fold nappes (F1), refolded by tight upright anticlinal and synclinal folds (F2), with axes plunging shallowly alternately to the northeast and to the southwest. Within the migmatites, however, this earlier structure is disrupted. Large blocks with coherent stratigraphy, whose lithologies resist migmatization, are isolated within highly mobile migmatites. This disruption of structure within the migmatites has resulted in a skewing of F1 fold axes. Juxtaposed against the migmatites, the unmigmatized rocks have a very different structural style. These rocks appear to be locally in faulted contact with the migmatites, and there is a dominant set of F3 folds plunging to the west along this contact.

Metamorphism. Billings (1941), recognized that the Mt. Washington area lies within the sillimanite metamorphic zone and distinguished seven subzones of metamorphic rocks on the basis of physical appearance and mineral composition. Three major stages of metamorphism correlated with orogenic stages are defined by Billings (1941).
The first stage is marked by andalusite, and is believed to have occurred prior to minor folding based on the andalusite pseudomorphs and sillimanite, which partially replaces the andalusite, being folded with the strata. The second stage is recognized by sillimanite, staurolite, garnet, tourmaline, and much of the muscovite and biotite. The muscovite and biotite are generally undeformed, but do show some evidence of deformation outlasting recrystallization by locally broken crystals. The staurolite and tourmaline are considered late minerals because of their diverse orientations. The third stage is retrograde metamorphism characterized by chlorite and sericite alteration.

Henderson (1949), recognized four phases of metamorphism in the Crawford Notch quadrangle that he correlates with various stages of the orogeny. The first stage of metamorphism involved simple recrystallization of the sediments producing schists and gneisses with no migmatization, early in the orogenic cycle. The next stage involves movement of material, but no changes in the bulk composition. The products were schist and gneiss with abundant granitic material; essentially a process of migmatization. The calc-silicate granofels were not affected by this metamorphic differentiation. The granofels were broken into fragments but not altered. The third metamorphic stage is an intrusive stage in which continued and intensified metamorphic differentiation formed local melts. The final retrograde stage is characterized by chlorite and sericite alteration. This stage occurred after the orogeny in a response to a decrease in temperature and pressure.

Wall (1988), in a study of the Pinkham Notch and Mt. Madison area, reports that the migmatized Rangeley and Littleton Formations are within the upper sillimanite zone and experienced maximum temperatures of 600° to 630°C and maximum pressures of approximately 3.5 Kb. Upper staurolite zone and lower sillimanite zone are also present in the pelitic schists of the Mt. Madison area.

Four metamorphic events are reported by Wall (1988). M1 is an early, syn-kinematic, low grade metamorphism. Alignment of minerals to form a schistosity is a result of this stage. All ensuing metamorphic events appear to be static. M2 is a regional event that reached at least staurolite + andalusite + biotite grade, with evidence based on partial to complete muscovite pseudomorphs of andalusite and staurolite. M3 is a regional, prograde and possibly retrograde event that is responsible for producing the mapped pattern of the upper staurolite, lower sillimanite, and upper sillimanite metamorphic zones. M2 and M3 are separated by a cooling event. M4 is a retrograde event in which chlorite replaces biotite, staurolite porphyroblasts are rimmed with sericite and staurolite megacrysts are partially to completely replaced by sericite and chlorite. Sillimanite is replaced by sericite in select areas.

Allen undertook detailed studies of unmigmatized schist and migmatitic gneiss outcrops that straddle a migmatite front bordering one of the migmatite zones in the Pinkham Notch area. Geochemical data show that the schists represent the parent material of the migmatites. Petrologic studies show that there is a steep metamorphic and thermal gradient across the migmatite front and that the migmatization reactions involved partial melting driven by infiltrating fluids. Stable isotopic data confirm that the migmatites were open to infiltrating fluids. Allen concludes that infiltrating fluids are responsible for the migmatization, but the source of these fluids remains elusive.

Allen (1992 and Allen & Chamberlain, 1992) has suggested that buoyant ascent of granitic magmas through highly mobile mixes of partially melted migmatites and granitic magmas may have been the source of the infiltrating fluids that created the migmatization. Allen also suggests that the Central New Hampshire Anticlinorium or “dorsal zone” of Eusden et al. (1987) and Eusden & Lyons (1993) may provide a structural control on pluton migration, and therefore migmatization, through the crust.

One of the smaller bodies of granite is the Wildcat Granite, which Allen (1992) proposes is equivalent to the Spaulding Group of the New Hampshire Plutonic Series. Two phases of the granite include the “G” phase.
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(white, medium-grained, two mica granite) and the “R” phase (biotite-rich, coarse-grained, rusty weathering with calc-silicate pods). The “R” phase possibly originates from melted Rangeley Formation.

A larger, mappable pluton is the Peabody River stock which is a homogeneous, undeformed two-mica granite or quartz-monzonite of the Concord Group of the New Hampshire Plutonic Series (Billings and Fowler Billings, 1975). The Peabody River stock was emplaced post-tectonically (Allen, 1992).

Allen (1992) has suggested that these plutons migrated from deeper levels through the crust and were the source of infiltrating hot fluids that caused migmatization. High-grade deep-level granulites and migmatites exposed in Massachusetts could represent the source regions of the granitic magmas (Allen, 1992). The magmas then moved upwards through the crust and formed metamorphic hot spots seen in New Hampshire. The plutons were finally emplaced at shallow crustal levels in Maine. Contact aureoles around these plutons provide evidence that the magmas did not melt in place, but in fact, migrated from deeper crustal levels.

The youngest phase of magmatism that affected the region is the White Mountain Magma Series, which consists of Early to Middle Jurassic overlapping volcanic/plutonic centers of felsic magmatism restricted predominantly to a north-northwest trend in New Hampshire. This series is characterized by the White Mountain batholith that intruded the Paleozoic metamorphic and plutonic rocks. The principal rock types of the White Mountain Magma Series near the Presidential Range are the Conway Granite, Moat Volcanics, Mt. Osceola Granite, and the Hart Ledge Complex (Henderson et al., 1977).

RESULTS

Figure 1 shows the geologic map of the fieldtrip area. Cross sections are shown in Figures 2 and 3. A stratigraphic column and explanation of rocks units are shown in Figure 4. Field trip localities are shown in Figures 1 and 5. Table 1 and Figure 6 present the sequence of sedimentation, deformation, metamorphism, and plutonism.

Stratigraphy

Five metasedimentary formations have been recognized in the alpine zone (Figure 1). They constitute a conformable stratigraphy, which, from oldest to youngest, consists of the Silurian (?) Rangeley, Perry Mountain, Smalls Falls, and Madrid Formations, and the Devonian (?) Littleton Formation. The stratigraphy is unfossiliferous and assigned ages are based on lithologic correlations; thus, queries are given after the age assignments. Within the northern Presidential Range and Clay Klippe, only partial sections of the stratigraphy are preserved. The missing sections are cut out by a stratigraphic discontinuity interpreted to be a thrust fault. The Mount Washington area and southern Presidential Range are stratigraphically contiguous. A general description of the five formations and stratigraphic columns for each domain is given below (Figure 4).

Littleton Formation: The Littleton Formation consists of dark gray schists commonly with interbedded quartzite layers of varying thickness and abundance. Andalusite, generally completely pseudomorphed by muscovite, sillimanite, and sericite, is common in the schists forming lumps, approximately 1 to 3 cm in diameter, and elongate aggregates, from 1 to 15 cm in length, with rare relict cores of fresh pink andalusite and/or chiastolite crosses. Schistosity in the schists is well developed and is usually parallel to bedding. In F1 fold hinges, bedding and schistosity become oblique to each other. The quartzites are fine-grained, light gray, and granoblastic. Graded beds, reversed in grain size by high grade metamorphism, are common throughout the formation.

The Littleton Formation has been subdivided into fifteen different members and three submembers based on variations in bedding style of the schists and quartzites and any other lithologic peculiarities (Figures 1 and 4). Representative, typical lithologies used to subdivide the Littleton are, in no particular order, (1) massive schist; (2) rhythmically bedded, thin-bedded schist and quartzite, the couplet being approximately 3 to 10 cm in thickness; (3) well-bedded schist and quartzite with graded bedding preserved, and quartzites generally between 10 and 50 cm in
Figure 2. Cross section through Mt Clay

Table 1. Sequence of depositional, deformational, metamorphic, and plutonic events in the Presidential Range.
Figure 3. Cross section through the Chandler Ridge Dome.
Figure 4. Representative stratigraphic columns showing lithologic variations and names of the various members of both the Littleton and Rangeley Formations
thickness; (4) well-bedded schist and quartzite with graded bedding preserved, and quartzites generally between 50 and 100 cm in thickness; and (5) massive quartzites at least 1 meter, and up to several meters in thickness, with thin, up to 10 cm thick, interbeds of schist and occasional graded bedding. All contacts between the members and submembers of the Littleton are gradational. Stratigraphic order of the Littleton members is exceptionally well controlled by the graded bedding.

**Madrid Formation:** The Madrid Formation is a fine-grained, thinly laminated, granofels with well-defined alternating layers of biotite-rich, schistose granofels and calc-silicate-rich granofels. The individual layers of granofels are from 1 to 5 cm thick. No graded beds are found. The formation weathers to a dark greenish-gray, is characteristically broken into platy fragments giving it a flaggy appearance. Total thickness of the formation varies between 10 and 50 meters. The contact between the Littleton and Madrid is abrupt and marked by the first appearance of the granofels. Within the granofels are one or two predominately schistose horizons, up to 2 meters in thick, that resemble the Littleton.

**Smalls Falls Formation:** The Smalls Falls Formation is a well foliated schist with distinct red-brown rusty weathering. The formation has a dark gray to black fresh surface, is highly susceptible to weathering, as a result often poorly exposed, and breaks into platy fragments generally smaller then those of the Madrid. Quartzite makes up less than 5% of the unit, with layers up to 5 cm in thickness, that are generally weakly magnetic due to the presence of pyrrhotite. No graded beds are found. Total thickness of the formation varies between 10 and 50 meters. The contact between the Smalls Falls and Madrid Formations is gradational over about 10 meters, the rusty schist gives way to progressively less rusty weathering, schistose granofels, and then ultimately to non-rusty granofels of the Madrid.

**Perry Mountain Formation:** The Perry Mountain Formation is a dark gray schist with interbedded light gray to white quartzites that are commonly 4 to 10 cm in thickness. Quartzites make up 30 to 40% of the unit and can range up to 60 cm in thickness. The contact between the Smalls Falls and Perry Mountain is abrupt. The unit is discontinuous ranging between 0 and 75 m in thickness. It is only exposed in the Tuckerman Ravine and Boott Spur region.

**Rangeley Formation:** The Rangeley Formation is a gray migmatitic paragneiss with abundant calc-silicate lenses. Angular to sub-rounded quartz and/or feldspar segregations (clasts?) between 2 and 8 cm in diameter are common. Elongate, rectangular lenses (clasts?), .5 to 2 meters in length, of well-bedded calc-silicate granofels and ellipsoidal lenses (clasts?), 10 to 50 cm long, of concentrically mineralogically zoned calc-silicate granofels without bedding are common throughout the Rangeley. A few beds of schist and quartzite are sometimes preserved, having escaped migmatization, and in these rare locations, no graded bedding is found. In places, the gneiss is extensively injected by pegmatites, aplites, and granite. The calc-silicate lenses are most often aligned parallel to schistosity, but some are at slight angles or, in the extreme, perpendicular to schistosity. The Rangeley Formation has been subdivided into three different members and six submembers based on lithologic variations in the gneiss and subordinate units of rusty gneiss, rusty schist, calc-silicate granofels, and amphibolite. Because the Rangeley is devoid of graded beds throughout the Presidential Range, stratigraphic order is based on the uninterrupted juxtaposition of the younger Smalls Falls, Madrid, and Littleton Formations. Where this juxtaposition is not available, as is the case for the Clay Klippe, the internal stratigraphic order of the Rangeley is indeterminate. The contact with the Perry Mountain is gradational, marked by the first appearance of calc-silicate lenses. The contact with the Smalls Falls, when the Perry Mountain is missing, is abrupt.

**Structural Geology**

The alpine zone of the Presidential Range preserves three phases of ductile folding (F1, F2, and F3) and a single phase of thrust faulting (T1) (Table 1). The phases of folding and the thrust faulting are treated as discrete, distinct phases of deformation, each with a unique set of fabrics or geometric characteristics. This method of data presentation does not rule out the possibility that the phases of folding and faulting may represent a continuum of
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The sequence of deformations will be discussed from oldest to youngest.

**D1 deformation.** The first phase of deformation is characterized by reclined to recumbent, isoclinal, similar, cylindrical F1 folds. These folds deform bedding, S0. No earlier fabric (schistosity or cleavage) has been observed to be folded by the F1 structures. The facing direction of the beds in F1 hinge zones is consistently northeast, east, or, southeast. The vergence of these F1 structures was also toward the east. Only in the region between the four mile post on the Auto Road and the Nelson Crag Trail at treeline, are there west and southwest facing F1 hinge zones in the Tuckerman Ravine syncline and these have been reoriented by the second phase of folding, D2.

Abundant L1 pseudoandalusite lineations are associated with F1 folds. L1 lineations are not uniformly distributed throughout the study area, are restricted to the Littleton formation, and within the Littleton formation there exist members where L1 is well developed and others members where only lumpy, unaligned, pseudoandalusites are found. Whenever L1 lineations are observed together with F1 fold axes, the two fabrics are parallel. As such, the lineations are best classified as b-type lineations.

**D2 deformation.** The second phase of folding is based on the map pattern and systematic variations in the attitude of S0 and S1 fabrics. Only three mesoscopic F2 fold hinges have been observed in the study area. The general style of folding is characterized by open, moderately to gently southwest or south plunging, moderately to steeply inclined folds, without an axial planar foliation. Within the Clay Klippe macroscopic F2 anticlinal axial traces are shown. The most spectacular F2 macroscopic structure is a broad partial dome, the Chandler Ridge dome (Figure 3), that is approximately 2 kilometers in width, and defined by both the contacts of the Littleton members and the F1 fold axial traces.

**T1 thrusting.** The area including Mount Clay and Great Gulf is interpreted to be a klippe (the Clay klippe) composed of Rangeley formation gneisses, separated from the Littleton formation to the north and south by the Greenough Spring thrust fault (Figure 2). The bases for establishing this as a thrust fault and klippe are, (1) the F1 folds in the northern Presidential range and Mount Washington area are truncated at the boundary; (2) the F2 folds within the klippe are also truncated at the boundary; (3) the boundary separates different metamorphic grades; sillimanite zone metamorphism in the northern Presidential range and Mount Washington regions, and migmatites, extensively injected by two-mica granites, within the klippe; (4) as the boundary is approached the pseudoandalusites in the Littleton formation become flattened; and (5) there is a significant stratigraphic discontinuity across the boundary; the Smalls Falls and Madrid formations are cut out.

**D3 deformation.** D3 deformation is the most common and abundant phase of folding seen in the alpine zone. F3 folds are characterized as mesoscopic, moderately inclined to overturned, generally gently but often moderately plunging, asymmetric F3 folds. The F3 folds have no, or only a weakly developed, axial planar S3 cleavage. F3 folding is not uniformly distributed throughout the study area. All areas show some F3 folding, but in places it is distinctly more pervasive than in others. This distribution appears to be related to the propensity of a particular rock to develop F3 folds, and not to heterogeneous distribution of strain related to D3.

The typical shape of an F3 fold consists of long, moderately dipping west limbs and short, steeply dipping east limbs. In some instances, the east limbs are overturned. This asymmetry gives a west over east sense of rotation, which we interpret to be related to east tectonic vergence of these folds. Pegmatite and alpine veins, dikes and sills seen throughout the alpine zone are folded by F3.

**Metamorphism and Plutonism**

**Metamorphism:** Five pulses of metamorphism, M1 through M5, have been recognized in the Alpine Zone (Table 1). M1 is characterized by aligned andalusite and pseudoandalusite and was syn-D1 based on the observation that the majority of L1 andalusites are parallel to F1 fold axes. M2 is characterized by sillimanite zone metamorphism in the Littleton schists and migmatization in the Rangeley. M2 occurred during the later part of D1 nappe-stage folding.
based on the observation that M2 fibrolitic sillimanite, early muscovite, and early biotite define the schistosity that is parallel to the S1 axial plane of the F1 nappes.

The Alpine Zone migmatites record two different types of sharp and abrupt contacts with the surrounding schists. The first type is "tectonic" and represented by the Greenough Spring Thrust fault. This discontinuity not only offsets the stratigraphy and early D1 and D2 folds, but also coincides exactly with the metamorphic transition between schists outside of the klippe and gneisses within. As the Greenough Spring thrust fault is approached the grade of metamorphism in the younger, structurally deeper schists, increases into the upper sillimanite zone and the pseudo-andalusites become flattened. Within the Clay Klippe all of the rocks have been extensively migmatized and injected with small apophyses of granite and pegmatite. Metamorphic grade is in the sillimanite zone with potassium feldspar-absent melting. The emplacement of the hot klippe apparently initiated prograde metamorphism in the cooler lower plate in contact with the Greenough Spring thrust.

The second type of migmatitic contact is "thermal" and observed in the southern Presidential Range where pelitic schists and quartzites, calc-silicate granofels, and rusty schists of the Littleton, Madrid, and Smalls Falls Formations respectively, give way to migmatites of the Rangeley Formation. There is no stratigraphic or structural discontinuity associated with this metamorphic transition, which is sharp and abrupt, occurring directly at the contact with the Rangeley. The migmatites themselves are characterized by sillimanite zone metamorphism with potassium feldspar-absent melting. This metamorphic transition seems less likely due to tectonism but rather more to compositional differences between the formations allowing the Rangeley to melt at lower temperatures.

M3 and M4 are contact metamorphic events both reaching staurolite grade in the Littleton schists. M3 occurred prior to D3 folding based on the observation that the early granites related to this phase of metamorphism are folded by F3 folds. 40Ar/39Ar mineral ages from late, coarse-grained, M3 muscovite were used to determine exhumation rates in the vicinity of the Auto Road (Eusden and Lux, 1995). The muscovite ages increase progressively from the bottom to the top of the mountain, the oldest being 304 ± 3 Ma and the youngest 274 ± 3 Ma. The age and elevation data yield an exhumation rate of 0.04 mm/yr. The exhumation rate indicates that (1) the Middle Pennsylvanian through Early Permian was a period of very slow exhumation; (2) rapid uplift associated with the Acadian orogeny must have terminated by at least 305 Ma; (3) sometime after 274 Ma, renewed uplift created the present topography of the Mount Washington massif; and (4) the relief method can be used successfully in older, more deeply eroded orogenic belts, if the relief is adequate and cooling is slow. M4 metamorphism occurred after D3 deformation and is related to the latest stage of post-tectonic granite intrusions. M5 is a retrograde metamorphism producing spotty occurrences of chlorite and/or sericite alteration in the schists and gneisses.

Plutonism: The plutonic history of the Alpine Zone is well linked to the structural and metamorphic history. Early plutonism is characterized by sills, veins, and plutons of two-mica granite, best exposed in the southeast portion of the Alpine Zone. M3 contact metamorphism was associated with the intrusion of these granites. These granites are subsequently deformed by F3 folds. Since these plutons do not have S1 foliation and are not involved with F1 folds (as the migmatites are), the processes of migmatization and plutonism in the Alpine Zone were apparently separate in space and time. Late stage post-tectonic granites are very rare and restricted to small apophyses in the northern Presidential Range where widespread, late M4 staurolite metamorphism is found. We are still looking for the larger pluton that caused all of this static staurolite zone metamorphism in the northern Presidentials!

DISCUSSION AND SUMMARY

Stratigraphic correlations and implications: The metasedimentary rocks of the Presidential Range have been correlated to the Silurian and Devonian Central Maine Terrane cover sequence. The maximum estimate of the total thickness of the Silurian and Devonian metasedimentary rocks in the alpine zone is in the range of 3,500 ± 500 meters (Figure 4). The Rangeley is the thickest formation at approximately 2,000 meters, but must be even thicker as the base was never observed. The estimated total thickness of the Perry Mountain, Smalls Falls, and Madrid Formations is 175 meters. The Littleton Formation is approximately 1,500 meters thick but also must be thicker as
the top was never seen. Based on estimates of the Silurian marine basin geometry by Moench and Pankiwskyj 
(1988), the thicknesses reported above would place the stratigraphic section of the Presidential range slightly seaward 
of the Silurian tectonic hinge.

The calc-silicate lenses in the Rangeley can be broadly characterized as mappable units of rock fragments 
enveloped by a mudstone-rich rock which has been metamorphosed into a gneiss. We interpret all of the Rangeley 
mapped in the Presidential Range to be a metamorphosed, variably migmatized, olistostromal melange.

Structural Geology: The sequences of ductile structures observed in the Presidential range more closely matched the 
structural models of Eusden and Lyons (1993) and Robinson et al. (1991) in New Hampshire and Massachusetts, and 
not the structural model of Moench and Pankiwskyj (1988) in western Maine. Apparently structures related to the 
emplacement of Devonian plutons (Moench and Pankiwskyj, 1988) do not extend west into the Presidential Range. 
The transition between the regime of nappe-dominated deformation and deformation caused by plutons must be east 
of the Presidential Range in the vicinity of the Maine-New Hampshire border.

D1: The interpreted east vergence of D1 nappes suggests that: 1) the dorsal zone model (Eusden and Lyons, 1993) 
should be restricted to central New Hampshire and the Central New Hampshire anticlinorium dies out south of the 
White Mountain batholith; 2) the model calling for west directed thrust-nappes, that are backfolded and domed 
adjacent to the Bronson Hill Anticlinorium (Robinson et al., 1991) should be restricted to the southwestern portion 
of the Central Maine Terrane; and 3) a revised regional model for Acadian deformations is needed to account for the 
D1 deformation in the Presidential range as well the structural transitions along the length of the Central Maine 
Terrane.

D2: In comparison to the regional models of deformation in the Central Maine Terrane, the second phase of 
deformation in the Presidential Range is quite unusual. It is poorly represented mesoscopically, yet is significantly 
affects the macroscopic map pattern, but only in random areas (the Clay Klippe and Chandler Ridge Dome). 
Furthermore, there are no D2 fabrics, lineations or foliations, developed in the rocks. We believe that these 
structures may be related to the collapse of the nappe pile after D1 deformation and peak metamorphism.

D3: The final phase of ductile folding in the Presidential Range, D3, is similar to most final phases of folding 
described for the Central Maine Terrane. These folds developed during the waning stages of regional metamorphism 
when P-T conditions were such that only a weak axial planar cleavage developed. Since these folds also deform two­mica granite sills in the Range, they postdate the earliest phases of granite plutonism.

Metamorphism. The metamorphism observed in the Alpine Zone occupies the transition between the metamorphism 
in Massachusetts, at the southern end of the Acadian metamorphic high, which is characterized by very high grade, 
syn-kinematic, deep-level gneisses and granulites (Tracy & Robinson, 1980; Robinson et al., 1989) and the 
metamorphism in western Maine, at the northern end of the Acadian metamorphic high, which is characterized by a 
shallow-level, high-grade, post-kinematic, metamorphism that is thought to be the result of overlapping contact 
aureoles around Acadian plutons (Guidotti, 1989; Lux et al., 1986). M1 and M2 in the Alpine Zone are more 
typical of the styles of metamorphism seen in central and southern New Hampshire, whereas M3 and M4 are more
typical of the metamorphism seen in Maine. It is likely that these metamorphic differences are attributed to systematic variations in crustal depths along the strike of the Acadian orogen, where deeper levels are now exposed in southern New England, with shallower levels to the north (e.g., Chamberlain & Robinson, 1989). The Presidential Range area appears positioned near the "top" of the region of syn-kinematic, high-grade metamorphism in New Hampshire, at the transition to the shallow-level, post-kinematic contact metamorphism in Maine.

REFERENCES


Guthrie, G.D. and Burnham, C.W., 1985, Petrology and origin of calc-silicate bodies from the Rangeley Formation,
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of Acadian Metamorphism with Depth in the Central Acadian High, New England, Contribution Number 63. Amherst: Department of Geology and Geography, University of Massachusetts, 69–140.


ROAD LOG

Time and place. Friday, September 26, 7:30 AM at the base of the Auto Road in the parking lot west of Route 16 and south of the entrance to the Auto Road. The base of the Auto Road is about 5 miles south of Gorham, N.H. A copy of Bradford Washburn’s topographic map of the Presidential Range (Washburn 1988) would be very helpful for locating the STOPS described below. STOPS are located on both Figures 1 and 4. Due to the fragile nature of the alpine ecosystem, please walk on trails or rocks where possible.

Warning. Expect extremely cold and miserable weather. Be prepared for below freezing temperatures, snow, sleet, hail, horizontal rain, and winds in excess of 50 mph. Bring warm clothes for winter-like conditions. Bring your lunch and water; there will be absolutely no chance to pick up anything once the trip begins. Bathrooms facilities are of the most primitive sort to be found. Bring hiking boots; there will be several strenuous hikes with considerable elevation loss and gain (+/- 1,000 feet elevation) of approximately 1 to 2 miles in length. We will drive up the Auto Road to the top of Mt. Washington and work our way down the mountain during the day. Vehicles must be consolidated. The folks at the Mt. Washington Auto Road have kindly agreed to charge us only $ 15 per vehicle. Vans are permitted up the road but only with seven (7) people, including driver.

MILEAGE

0.0  Base Station. Drive from the base of the Auto Road to the summit of Mt. Washington

8.0  Lower parking lot (elev. 6,200 ft.) of Mt Washington Summit, labelled P1 on Figure 5. Park vehicles and walk to STOP 1A-B

STOP 1A. In blasted outcrop along the NW side of the road just below parking lot, inverted graded bedding of Huntington Ravine member of Littleton Formation within inverted limb of the Mt. Washington Nappe, small sills of pegmatite as well.

Walk down the road about .1 mile just below the parking lot with five oil tanks on the west side of the road at about elevation 6,100 ft.

STOP 1B. On the east side of (downhill side) the road are blocks and outcrops of schists and thin quartzites of the Huntington Ravine member of the Littleton Formation with excellent L1 pseudoandalusite
lineations. Where aligned, the L1 orientation is parallel to the F1 fold axes throughout the Presidential Range.

Return to vehicles and drive down the Auto Road.

8.6 Park vehicles in lot on west side of road which is just above the 6,000 ft elevation post, labelled P2 on Figure 5. STOP 2 is a 2.5 mile walking loop to several outcrops.

STOP 2A. Walk .1 mi. east and slightly uphill to Ball Crag (elev 6,112 ft.) and examine inverted graded beds of the Huntington Ravine member of the Littleton Formation with well developed F3 folds and poorly developed S3 axial plane cleavage.

Walk back to the parking lot with the vehicles and walk .1 mi. west to the intersection of the Great Gulf and Gulfside Trails. Follow the Gulfside Trail WNW paralleling the Cog Railroad, passing the junction of the Westside Trail, to the Clay Col and junction of the Gulfside and Clay Loop trails, about .9 mi.

STOP 2B. In the Col the geology passes from schists of the Bigelow Lawn member of the Littleton Formation to the SE to gneisses and abundant pegmatites of the Mt. Clay member of the Rangeley Formation to the NW. The Madrid, Smalls Falls, and Perry Mountain Formations are cut out here. The contact between the Rangeley and Littleton is not exposed, but you can get within about 20 meters of it; look for the last outcrop of schist and the first outcrop of gneiss with calc-silicate lenses. This contact we have interpreted as the Greenough Spring Thrust Fault.

Proceed N along and up the Clay Loop Trail for about .1 mi.

STOP 2C. In the trail and just to the W on the first cliff up from the Col, are lichen-free outcrops of the Mt. Clay member of the Rangeley Formation gneisses with abundant quartz segregations (clasts?) and rare, oval, concentrically zoned, calc-silicate lenses (clasts?). Proceed up the first steep pitch on the Clay Loop Trail another .05 mi., which is halfway up the first or S summit of Clay. On the E side of the trail are exposed meter-long, bedded clasts of calc-silicate with bedding truncated by the gneissic matrix. We have interpreted this type of Rangeley to be a metamorphosed olistostromal melange.

Proceed over the first summit of Clay, which has a small saddle occupied by a buried basalt dike, to the 200 yard wide grassy col between the two principal summits of Clay, about .3 mi. from STOP 2C. Descend .1 mi. to the east towards Great Gulf down the progressively steepening gully from the col to about elevation 5,400 ft.

STOP 2D. On the south side at the top of the steepest and narrowest part of the gully is an outcrop of the Mt. Clay calc-silicate granofels member of the Rangeley Formation. This was not previously mapped by the Billings. We have interpreted it to be a 100 m long, 30 m wide block of calc-silicate within the Mt. Clay member of the Rangeley Formation.

Return to vehicles along Clay Loop and Gulfside Trails, about 1.3 mi., and drive down the Auto Road.

9.5 Park at the lot on the NW side of the road at elevation 5,700 ft.; the relatively flat area of the so-called “Cow Pasture”, labelled P3 on Figure 5. STOP 3 is a 1 mile walking loop to several outcrops.

Walk to the junction of the Nelson Crag and Huntington Ravine Trails and proceed up the Nelson Crag Trail about .1 mi. to about elevation 5,800 ft.; this would be between the 14th and 15th cairns on the Nelson Crag Trail from the junction. Turn off the trail to the south, parallelising the slope and contours for about .1 mi. to prominent outcrops marked by a small cairn.
STOP 3A. F1 hinge zone exposures of the Tuckerman Ravine Syncline with thick bedded quartzites and thin schists of the Alpine Garden member of the Littleton Formation. S0, bedding, and S1, axial planar schistosity, are perpendicular here. Graded beds in the hinge zone suggest northeasterly facing directions for the F1 folds.

“Rockwack” downhill, southeast, for about .2 mi to the headwall of Huntington Ravine and the junction of the Huntington Ravine and Alpine Garden Trails.

STOP 3B. Pinnacle Gully dike and upright schist and thick micaceous quartzites of the Huntington Ravine member of the Littleton Formation.

Follow the Alpine Garden trail .2 mi northeast to junction with the Nelson Crag Trail.

STOP 3C. Exposed is a rare outcrop of an F2 warping fold about 15 yards southeast of the big cairn at the junction. This is one of only a handful of mesoscopic F2 folds seen in the Presidential Range and happens to be along the crest of the most significant macroscopic F2 structure, the Chandler Ridge Dome.

Follow the Nelson Crag trail about .1 mi. uphill to the unnamed crag at elevation 5,735.

STOP 3D. On the north and northeast flanks of this crag are upright beds of the Cow Pasture member of the Littleton Formation. S0 and S1 are now parallel here as the F1 hinge zone of the Tuckerman Ravine syncline is slightly above us.

Proceed .1 mi. to the junction with the Huntington Ravine Trail and then back to the parking lot with the vehicles. Drive down the Auto Road.

10.7 Park in the vicinity of Cragway Spring (elev. 4,800 ft.) in the small parking lots above and just at the hairpin turn, labelled P4 on Figure 5. STOP 4 is a 1.5 mile walking loop to several locations.

Hike down the Nelson Crag Trail .6 km to treeline at approximately 4,300 ft. elev.

STOP 4 A. Outcrops of the Great Gulf member of the Littleton Formation. Exposed here are schists with 10 cm thick quartzite interbeds. Bedding, S0, and schistosity, S1, are not parallel indicating that we are in the F1 hinge zone of the Horn Nappe. S1 schistosity is refracted through the inverted bedding defined by the quartzites. Fresh, pink andalusite and andalusite rimmed by fibrolitic sillimanite is abundant. A small cairn slightly north and a bit downhill from the trail marks an F1 fold that has been reoriented by the F2 Chandler Ridge Dome.

Proceed back up the Nelson Crag Trail about .2 mi through the first patch of scrub spruce and into the next higher treeless area. Head off the trail to the south about .05 mi to the prominent outcrops.

STOP 4B. Exposed here are the predominately massive schists of the Cow Pasture member of the Littleton. L1 lineations are folded by abundant F3 folds which define the topography here. Bedding and S1 schistosity are again parallel as we are now above the Horn Nappe F1 hinge zone.

Proceed back up the Nelson Crag Trail about .1 mi through another patch of scrub spruce and to the higher treeless area that includes the Auto Road at Cragway Spring. Head off the trail to the south about .05 mi.

STOP 4C. Exposed are schists and thin quartzites of the Huntington Ravine member. Well developed F3 folds fold thin aplite sills here. These small granitic apophyses are the earliest phase of granitic intrusion.
and are pre-F3 in age. Some of the larger granite plutons of this generation impart local staurolite grade metamorphism (M3) to the schists.

Return to the vehicles examine the road cuts and outcrops on the “inside” of the hairpin turn.

STOP 4D. Exposed are schists and thin quartzites of the Huntington Ravine member. Upright graded beds are common. Pseudoandalusite is replaced by fibrolitic sillimanite in the core with coarse-grained, M3 muscovite and staurolite in the rim. A great place to see this texture is a 1 m diameter lichen-free zone of bedrock marked by a small patch of concrete (4 cm diam) with a small metal pin (1 cm diam) embedded in it. F3 folds are everywhere.

Return to vehicles and drive down the Auto Road.

11.7 Park in the lot at elevation 4,200 ft. just below the junction with the Winter Cutoff Road, labelled P5 on Figure 5. STOP 5 is a short walking loop of about 1 mile.

Walk back up the Auto Road to the Winter Cutoff road and proceed up that about .1 mi to elev. 4,400 ft.

STOP 5 A. Marked by two small cairns are thick schists and 10-20 cm thick quartzites of the Great Gulf member of the Littleton. These outcrops show the exact hinge location for the Horn Nappe as bedding and schistosity are perpendicular. S1 is again refracted through the quartzites.

Walk about .2 mi up the ridge between the Winter Cutoff and Auto road to about elevation 4,450 and then bushwack down to the Auto Road. Walk down the road about .1 mi.

STOP 5B. Nice exposure of a rare F1 mesoscopic fold hinge in the outcrops on the west side of the road.

Walk about .1 mi down the road.

STOP 5C. Garnet-bearing granite sill exposed in a road outcrop. Would anyone like to date this for us? An age determination would nail down the timing of F3 folding and M3 contact metamorphism in the Range.

Return to vehicles and drive down the Auto Road.

11.9 Park in the lot on the east side of the road just above the 4,000 ft. elevation post, labelled P6 on Figure 5.

Walk down the road about .2 mi to just below the 4-mile post.

STOP 6. Examine the large outcrop of thick quartzites of the Oakes Gulf member of the Littleton. There is one enormous bed, 5 meters in thickness, which is wonderfully graded and inverted. Bedding and schistosity are not parallel, meaning we are still within an F1 hinge zone.

Return to vehicles and proceed down the Auto Road.

16.0 Base Station and END OF TRIP.
Figure 6. Schematic portrayal of the sequence of tectonic events in the alpine zone of Presidential Range.
A. Rangeley deposition; B. Rangeley disruption; C. Perry Mtn., Smalls Falls, Madrid, and Littleton deposition; D. D1 nappe stage folding; E. D2 collapse of nappe pile; F. T1 Greenough Spring thrust fault; G. D3 final pulse of Acadian folding. Patterns: random dashes - Bronson Hill basement; crosses - Avalonian basement; black - Kronos crust; dot with radial dashes in B. - triggering seismic event to disrupt Rangeley. All sections are oriented west (left) to east (right).