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STRATIGRAPHIC AND STRUCTURAL TRAVERSE OF MOUNT MORIAH AND THE WILD RIVER WILDERNESS AREA

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INTRODUCTION

East of Pinkham Notch and south of the Androscoggin River valley lies a large roadless area of the White Mountain National Forest containing the Carter–Moriah mountain range and the Wild River valley (Fig 1). Original mapping in this and adjacent areas was done by M. P. Billings and others in the 1940's and 1950's (1941, 1946, 1975, 1979), and Billings' work lays the foundation upon which all subsequent work in this area is based. Since the time of Billings' maps, however, further developments have been made in understanding the litho-stratigraphy of Siluro-Devonian metasedimentary rocks throughout New Hampshire and western Maine (Osberg et al., 1985; Moench & Pankiwskyj, 1988; Lyons et al., 1991). Attempts to apply the new lithostratigraphy to this region met with limited success, particularly in the migmatites that make up the Carter–Moriah and Wild River area (CM–WR; Hatch et al., 1983; Hatch & Moench, 1984; Hatch & Wall, 1986). The CM–WR area is shown on current maps as "undifferentiated sedimentary rocks in areas of extreme migmatization" (Osberg et al., 1985; Lyons et al., 1991).

The rocks are indeed migmatized, although in general the line bounding these "undifferentiated" rocks does not necessarily represent a "migmatite front" but rather the limits of easily mappable terrain (Hatch & Wall, 1986, page 146). Locally, however, sharp "migmatite fronts" can be clearly defined, separating un-migmatized schists from intensely migmatized gneisses of the same parent lithology. Associated with detailed studies of such a front in Pinkham Notch (trip C6 of this volume; Allen, 1992, 1996b, Allen et al, 2001), I have undertaken a third-generation mapping effort attempting to differentiate the stratigraphy and structure of the migmatites in the CM–WR area.

REGIONAL GEOLOGIC SETTING

The migmatites in the CM–WR region lie along the axis of the Central Maine Terrane (CMT, Fig. 2; Zen et al., 1986) and are central to the broad region of high grade metamorphism of the Acadian Orogen. To the west and north, the CMT abuts the Bronson Hill Anticlinorium, which represents a magmatic arc of Ordovician age with a thin cover of Silurian and Devonian sediments. The CMT is regarded as an eastward thickening sedimentary basin adjacent to the arc, filled with Silurian age shales, quartzites, and calcareous rocks deposited in a deep water anoxic environment, and topped by early Devonian volcanics and turbidites from an eastern source (Moench & Pankiwskyj, 1988). This basin, together with the Bronson Hill arc, was multiply deformed and metamorphosed during large scale crustal thickening of the Acadian Orogeny.

The structure of the CMT consists of two major synclinoria separated by the Central New Hampshire Anticlinorium (Eusden, 1988). This anticlinorium acts as a "dorsal zone" from which originate west-vergent structures to the west and east-vergent structures to the east (Eusden, 1988). In fact, large scale west-vergent fold nappes carried high-grade rocks from the western CMT over the Bronson Hill Anticlinorium (Thompson et al., 1968; Chamberlain et al., 1988). The Central New Hampshire Anticlinorium is marked not only by exposures of the oldest rocks, but is also the locus of anomalous metamorphic "hot spots" (Fig. 2; Chamberlain & Lyons, 1983; Chamberlain & Rumble, 1988) and migmatite zones (Fig. 2; Wilson, 1969; Billings & Fowler-Billings, 1975; Englund, 1976; Eusden, 1988).

The plutonic rocks associated with the Acadian metamorphic high, and the Central New Hampshire Anticlinorium structure of the CMT, belong to the New Hampshire Plutonic Series. Petrologic and geochemical studies suggest that they are anatectic crustal melts (Duke, 1978; Clark & Lyons, 1986; Lathrop, et al., 1994). The oldest of three generations (400 to 390 Ma; Lyons & Livingston, 1977; Barreiro & Aleinikoff, 1985) are the synmetamorphic and syntectonic Kinsman, Bethlehem, and Spaulding groups. These are large, shallow sheet-like bodies that are intimately involved with Acadian nappe structures (Nielson et al., 1976; Thompson et al., 1968), and have isograds mapped across them (Chamberlain & Lyons, 1983). At about 380 Ma, post-tectonic plutonism resulted in abundant small bodies of two-mica granites known as the Concord group (Lyons et al., 1982; Harrison et al., 1987). The last group of the New Hampshire Plutonic Series yield ages of about 320 Ma (Lyons et al., 1991; Osberg et al., 1985), volumetrically significant only in Maine, causing late thermal metamorphism.

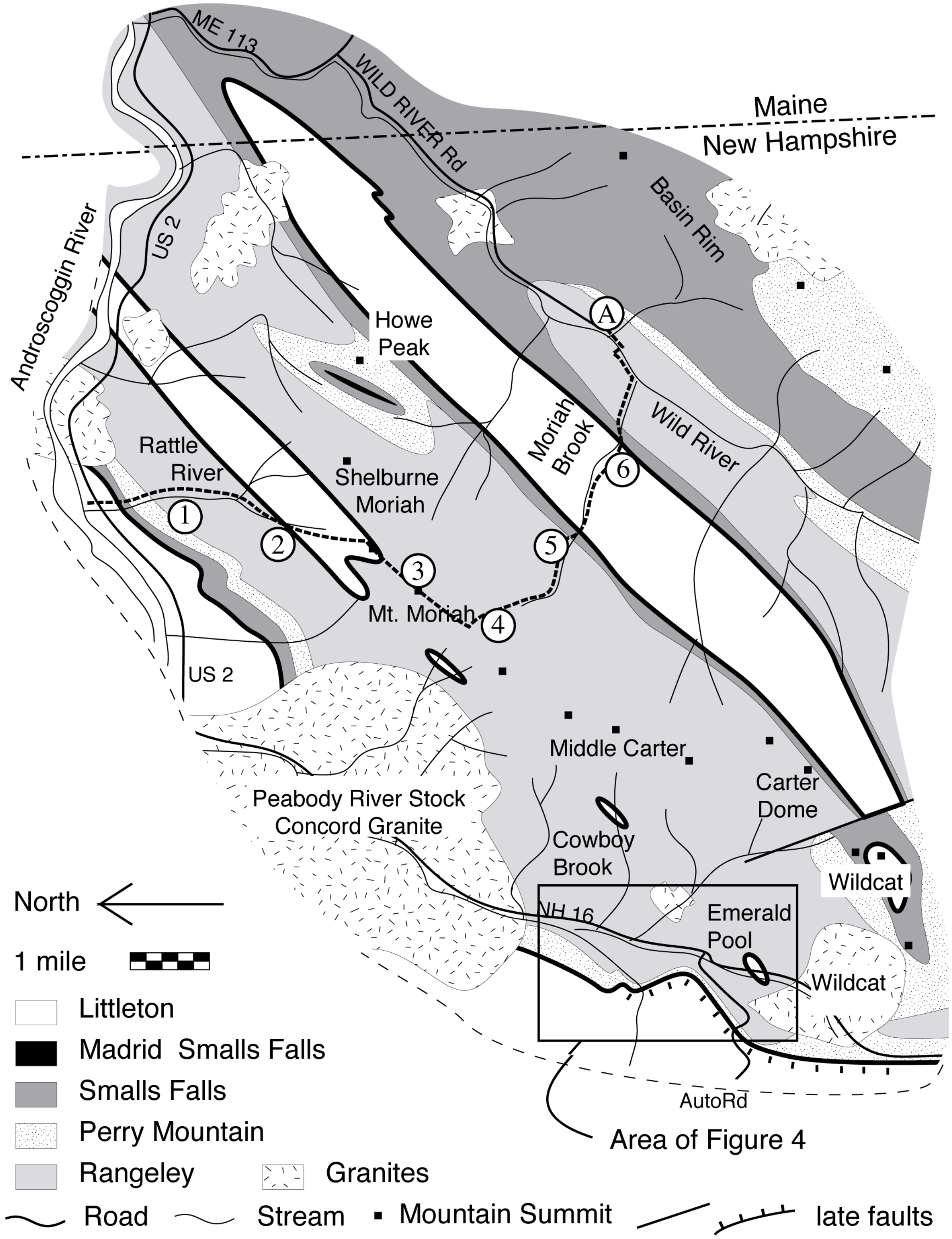

Figure 1: Geologic map of the Carter–Moriah and Wild River area, from this study and from Billings & Fowler-Billings (1975); Hatch & Wall (1986); and Hatch & Moench (1984). There are numerous granites and pegmatites too small to show. Note direction of north arrow. Field trip stops are indicated by the circled numbers.

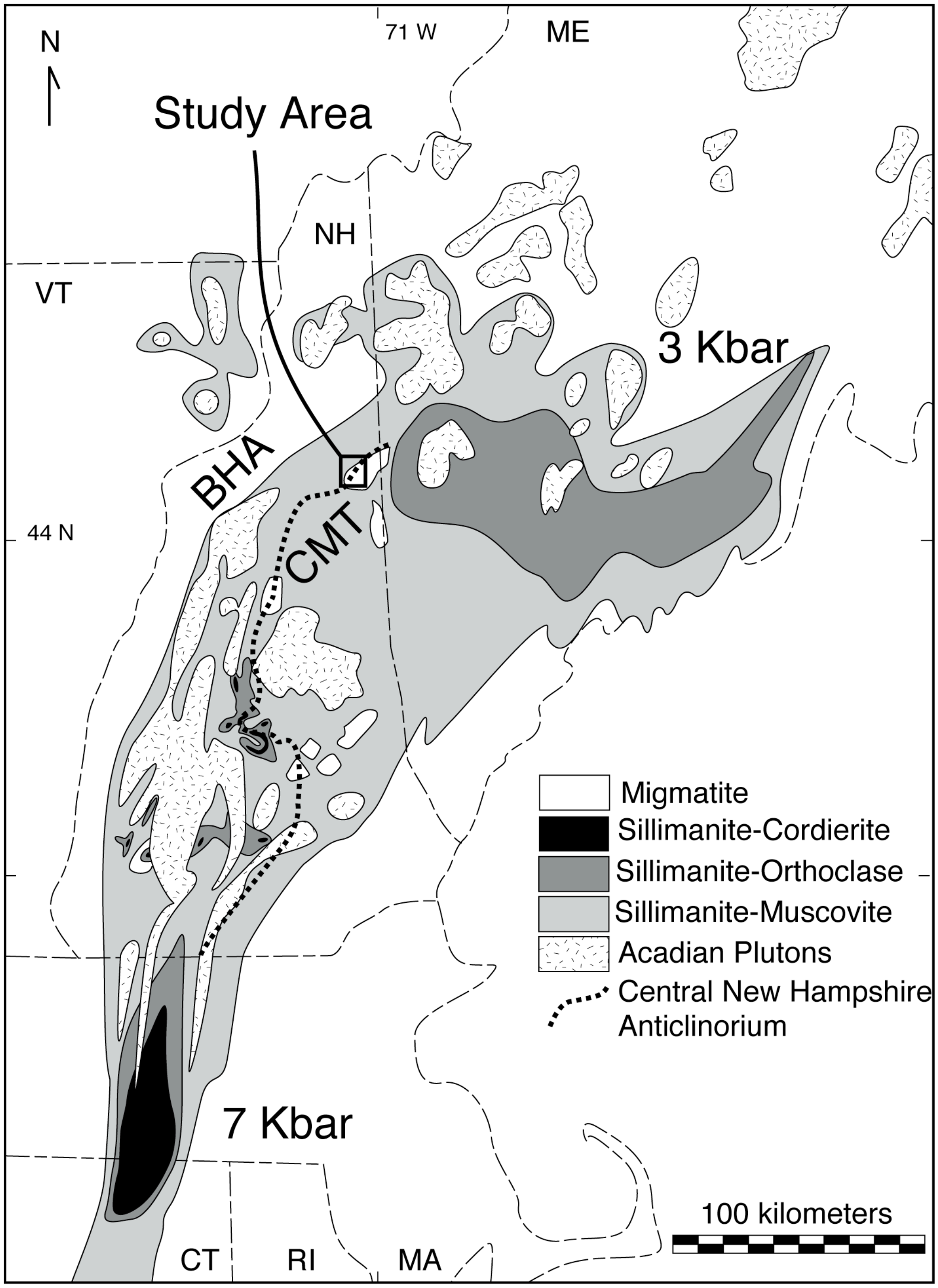

Figure 2: Map of the Acadian metamorphic high in New England, showing shallow level plutonism in Maine, "hot spots" and migmatite zones (schematic) and the Central New Hampshire Anticlinorium in New Hampshire, and deep-level, high-grade gneiss in Massachusetts. CMT labels the axis of the Central Maine Terrane; and BHA the Bronson Hill Anticlinorium. After Chamberlain & Robinson (1989).

METASEDIMENTARY ROCKS

Although Billings & others (1941, 1946, 1975, 1979) did not recognize the stratigraphic sequence as it is now understood (Hatch et al., 1983), they did recognize some of the important lithologies upon which the modern litho-stratigraphic sequence is based. Thus their maps provide a useful starting point for work in this region. All of the rocks previously assigned to the Devonian Littleton Formation (Billings & Fowler-Billings, 1975), are here subdivided into a Siluro-Devonian lithostratigraphy similar to that now recognized elsewhere in New Hampshire and western Maine, based on correlations made by Hatch et al., (1983). Units include the Silurian Rangeley, Perry Mountain, Smalls Falls, and Madrid Formations, as well as the Devonian Littleton Formation. The Perry Mountain Formation had not previously been recognized in much of this area (Hatch et al., 1983; Hatch & Wall, 1986). The metamorphism of these rocks is discussed elsewhere (trip C6 of this volume; Allen 1992, 1996b).

Rangeley Formation

The Rangeley Formation is the oldest and the most abundant rock in the CM–WR area. The Rangeley consists of gray and rusty orange weathering coarse-grained pelitic schists and migmatite gneisses, with minor interlayered quartzites, and abundant, distinctive calc-silicate pods. These rocks are generally biotite schists or gneisses, containing quartz, biotite, muscovite, albitic plagioclase, garnet and sillimanite. An important accessory mineral is pyrrhotite, whose presence is responsible for the orangish weathering color these rocks often have. It is the Rangeley that is most often migmatized in this area, although the intensity of migmatization is highly variable. I have not further subdivided the Rangeley in this area, as has been done elsewhere (Lyons et al., 1991; Eusden, 1988; Moench & Boudette, 1970).

Original bedding can occasionally be recognized where well-defined beds of contrasting composition occur. Graded beds or other topping indicators have not been observed. More often, original bedding cannot be recognized, although the schistose foliation and migmatitic layering are generally bedding-parallel. Many weathered outcrops present a massive appearance, however, and the fabric of the rock is best seen on water-washed polished surfaces. The character of the foliation in these rocks is very rough, as the rock is very coarse grained, often contains large, un-oriented muscovite spangles, and is typically migmatitic. Billings & Fowler-Billings (1975) described three types of migmatite fabrics: “layered”, “podded”, and “wispy.” These generally grade into one another, and are often highly distorted and incoherent—“swirly.”

The calc-silicate pods often occur in groups along bedding planes—often associated with more quartz-rich beds—or in clusters within larger quartz-rich pods or blocks. The pods resist migmatization, and in the migmatites they usually remain parallel to the foliation and gneissosity, although occasionally they are at odd angles to it. The pods are usually concentrically zoned in composition and rimmed by a weathered-out moat, while the cores stand up in relief above the surface of the surrounding schist or gneiss. These pods are thought to be metamorphosed calcareous concretions (Billings & Fowler-Billings, 1975), although some may be rip-up clasts or blocks of reef material carried from the shelf into the deep basin (Guthrie, 1984; Guthrie & Burnham, 1985), during the rapid sedimentation characteristic of the Rangeley (Moench, 1970). Elsewhere, calc-silicate pods in the Rangeley have been interpreted as boudins of once continuous calcareous sandy beds or lenses (Eusden, 1988). The appearance of “pods within block” features suggests that they are concretions or clasts and not boudins.

Also distinctive in the Rangeley of this area are occasional exotic quartz, quartzite, or granitoid pebbles and cobbles, usually as isolated individuals rather than in recognizable conglomeratic horizons, but clearly of sedimentary origin. Zones of matrix-supported polymictic conglomerate are observed in the Rangeley in other regions of Maine and New Hampshire (e.g., Moench & Pankiwskyj, 1988; Allen, 1984).

In addition to the calc-silicate pods and the cobbles, larger (1 to 5 meter) exotic blocks have also been observed. Often the lithology of these blocks is suggestive of the rock units that overlie the Rangeley. These blocks may have an origin similar to the rip-up clast model proposed for the calc-silicate pods. These features may represent a sedimentary or olistostromal mélange indicative of rapid, sometimes chaotic, sedimentation in a submarine fan environment (Eusden et al., 1996). Moench (1970) discusses extensive pre-metamorphic deformation in the Rangeley Formation due to extremely rapid sedimentation and the build-up of high fluid pore pressures.

Perry Mountain Formation

The Perry Mountain Formation consists of interbedded gray quartzites and schists, often bearing ptygmaticly folded coticules, and occasionally, calc-silicate pods. The quartzite and schist interbeds range from 1–2 cm to 5–10 cm in thickness, and occasionally quartzite dominates. The contacts between the quartzite and schist interbeds occasionally show grading, but are more often sharp. As in the Rangeley, the calc-silicate pods occur strung out along bedding planes, although the "pods within block" features are unique to the Rangeley. In the Perry Mountain, some of the pods are clearly boudins. Locally the schist layers have well developed sillimanite nodules, pseudomorphs after andalusite. The Perry Mountain is rarely migmatized in this area.

Most of the rocks shown as Perry Mountain Formation in Fig. 1 were not recognized as such by Hatch & others (1983, 1984, 1986), but instead were assigned to the Rangeley Formation. I have assigned these rocks to the Perry Mountain, however, because of the abundance of quartzites and the nature of the bedding, and because they appear in the proper sequence between Rangeley schists and migmatites and Madrid or Smalls Falls rocks.

Smalls Falls Formation

The Smalls Falls Formation is a highly graphitic and sulfidic schist with sulfidic micaceous quartzites. These rocks often weather a dull brown color to very dark rusty-red. The intense weathering due to the abundant pyrrhotite makes the rock very crumbly—as a result outcrops are not very resistant to erosion. The micaceous quartzite beds often breakdown to a characteristic gritty sand. These rocks are often finely laminated—flaggy—with bedding less than 1 cm thick. In many places, however, any semblance of bedding has been destroyed due to incompetent structural behavior.

Madrid Formation

The Madrid Formation consists of green calc-silicates, and fine grained plagioclase-biotite-quartz "salt & pepper" granofels. These rocks are very well bedded, weathering to produce distinctive tabular blocks and slabs, from 2 to 10 cm thick. The Madrid is generally not well exposed, being less resistant to erosion than adjacent quartzites and sillimanite schists; and is fairly thin throughout the region (never more than a few tens of meters). The best exposures are found in streams, and are quite distinctive. The Boott Member of the Littleton Formation and other lime-silicate rocks as mapped by Billings & others (1941, 1946, 1975, 1979) are now thought to be the Madrid Formation (Hatch et al., 1983).

Littleton Formation

The Littleton Formation in this area is composed of aluminous schists with interbedded quartzites, with generally very good graded bedding. These rocks are generally silver-gray in color, and have distinctive, abundant sillimanite pseudomorphs after andalusite, often up to 5 cm long. Locally, these sillimanites define a strong lineation; elsewhere they may occur as "turkey tracks" on foliation planes. Rocks of the Littleton Formation that have been migmatized often have a very different texture or fabric from the migmatites of the Rangeley. This may be best described as a "stringy" or "sinewy" texture, as the migmatite leucosomes appear to define a lineation within the melanosome/mesosome matrix. The calc-silicate pods abundant in the Rangeley are absent from the Littleton Formation.

In the Moriah Brook section (Stop 6, Fig. 1) there are conglomeratic horizons in contact with exposures of the Madrid Formation. Sequentially, these conglomerates appear to belong to the lower Littleton Formation. These conglomeratic horizons may be similar to the "Wild Goose Grits" mapped within the Littleton by Eusden & others (1987; 1988) south of this region. These conglomerates, and the apparent local absence or extreme thinning of the Madrid Formation, may indicate a local unconformity at the base of the Littleton.

IGNEOUS ROCKS

Billings & Fowler-Billings (1975) mapped several igneous rock types in the CM–WR area, belonging to the Devonian New Hampshire Plutonic Series, and to the Jurassic–Cretaceous White Mountain Plutonic-Volcanic Series. Rocks of this later series consist of volcanic vent agglomerate and diabase dikes, of minor importance to this study. Plutonic rocks in this region occur in two main modes—as large, mappable plutons such as the Peabody River Stock (Fig. 1; Billings & Fowler-Billings, 1975), and as smaller, more heterogeneous granitic and pegmatitic bodies and dikes that occur throughout the migmatite zone. As Billings & Fowler-Billings note (1975, p. 64), these rocks are difficult to portray on the geologic map because of their small size, wide distribution, and intricate contact relationships.

A larger body of this latter type occupies an area of about 10 km2, extending from Pinkham Notch proper north nearly to Emerald Pool, and east almost to the summit of Wildcat Mountain (Fig. 1, Stop 9). It underlies the slopes of the Wildcat Mountain Ski Area, from which is derived the name I have assigned to this type of rock—the "Wildcat Granite." The Wildcat Granite can be described as granite only in generalities—there are clearly at least two different phases. One consists of medium grained, whitish-weathering clean two-mica granite (hereafter, the "G" phase, for granite). The second is much coarser grained, orangish-weathering granitoid (hereafter, the "R" phase, for the Rangeley Formation), also bearing both muscovite and biotite, but with much more abundant biotite than in the "G" phase. Within this second phase are abundant calc-silicate pods , identical to those found in the metasediments, rimmed by strong reaction zones. Textures and mineralogy of the "R" phase suggests that it may be formed from completely melted and recrystallized Rangeley schists. Both the "G" and "R" phases are extensively intermingled in a highly complex fashion. Wispy biotite-rich schlieren can be observed throughout. The contact between the granite and the surrounding migmatitic metasedimentary rocks is gradational—not a sharp intrusive contact. Similar occurrences of Wildcat-type granitoids are found throughout the migmatite zone, usually associated with pegmatites. The Wildcat Granite is very similar to some exposures of the Blackwater Pluton of the Spaulding Group of the New Hampshire Plutonic Series (Lyons, 1988; Duke, 1978). The Spaulding Group is considered to be late-tectonic, and has been dated at 392 ± 5 Ma (Lyons & Livingston, 1977). Beyond the explanation given above, it is interesting to speculate why Billings & Fowler-Billings (1975) might not have shown the Wildcat Granite on their map—granitization and the origin of granites were "hot topics" at the time they were doing this mapping (1950's).

Two generations of granitic pegmatite are observed throughout the migmatite zone. The first is generally gray in color and contains quartz, albitic plagioclase, muscovite, and spessartine garnet. These pegmatites are sometimes slightly foliated, and often have gradational contacts with adjacent migmatites. The second generation pegmatites, which cross-cut the earlier pegmatites and the granites, are white in color, and contain quartz, albitic plagioclase, muscovite ± potassium feldspar, and abundant tourmaline. These white pegmatites usually have sharp contacts with adjacent rocks. Both sets of pegmatites can be either cross-cutting or parallel to the structural trends in the adjacent rock—the second-generation pegmatites also cross-cut the Wildcat Granite. Eusden (1988) has also observed two stages of pegmatite similar to that described here, associated with another migmatite zone to the south.

The Peabody River Stock (Fig. 1) is a homogeneous and undeformed two-mica granite or quartz monzonite of the Concord Group of the New Hampshire Plutonic Series (Billings & Fowler-Billings, 1975). This homogeneous granite is distinct from the Wildcat Granite, which is extremely heterogeneous. It appears to have no relationship to the migmatization. In fact, at the migmatite front to be visited on trip C6 (this volume), the rocks become more intensely migmatized as one moves away from the Peabody River Stock. The Concord Group granites represent post-tectonic plutonism at about 380 Ma (Lyons et al., 1982; Harrison et al., 1987), which is consistent with the undeformed nature of the granites in Peabody River Stock, and suggests that its emplacement post-dates the migmatization, and the associated Wildcat Granite.

STRUCTURE

Billings & others (1941, 1946, 1975, 1979) undertook extensive detailed study of geologic structures in the CM–WR area and adjacent Mount Washington region. Their analysis of the small scale structures and structural fabrics is excellent, however their interpretation of the larger structure was limited by the poor stratigraphic control (Billings & Fowler-Billings, 1975, page 57).

The mesoscopic, measurable structural features observed in these rocks are dominated by a foliation or schistosity defined by the alignment of micas. This foliation is generally bedding-parallel. Where the rock is migmatized, the leucosomes often lie in planes parallel to this foliation. Many of the rocks also have a strong lineation that is the result of small scale crenulation folds of the primary foliation, crenulation and alignment of micas, and the alignment of minerals such as sillimanite. The mineral and crenulation lineations are parallel to the axes or hinge lines of minor folds. Rarely can a good determination of the orientation of the axial plane to these fold features be made. A spaced cleavage, that may be axial planar to F3 folds, is observed only along the western boundary of the migmatite zone.

Overall, the structural trend in the migmatite zone is similar to that shown on Billings & Fowler-Billings (1975) map of the Gorham 15' quadrangle, as rock units and planar features strike northeast throughout the migmatite zone (Fig. 1). This trend is due to tight upright anticlinal and synclinal folding. This folding is demonstrated by the

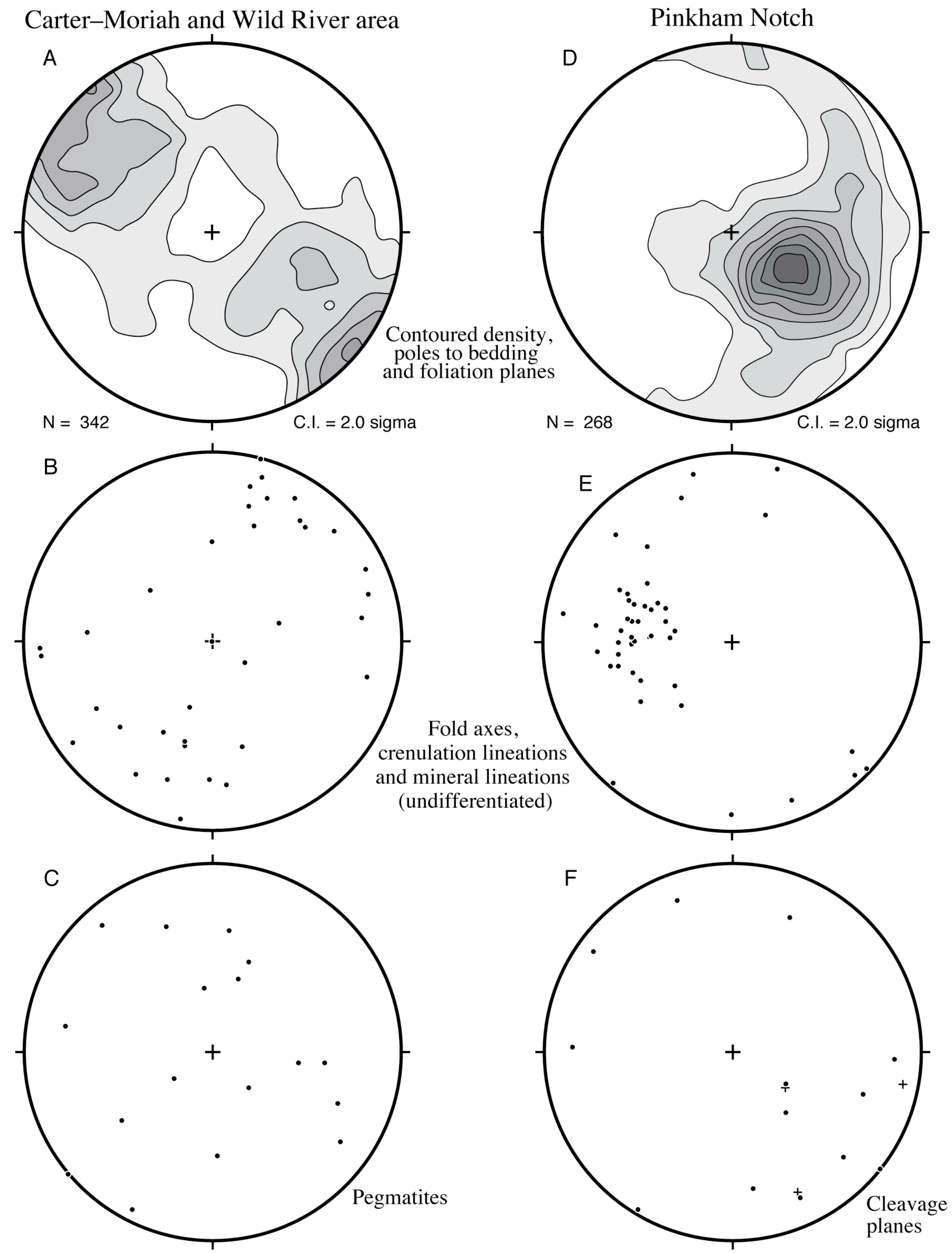

Figure 3: Equal Area stereonet projections of: Contoured density of poles to bedding and foliation planes from the Carter Moriah and Wild River area (A) and from Pinkham Notch (D); Fold axes, crenulation lineations and mineral lineations (undifferentiated) from the Carter–Moriah and Wild River area (B) and Pinkham Notch (E); poles to planar pegmatite contacts from the Carter–Moriah and Wild River area (C); and Poles to spaced cleavage (•) and F3 axial planar cleavage (+) from Pinkham Notch (F).

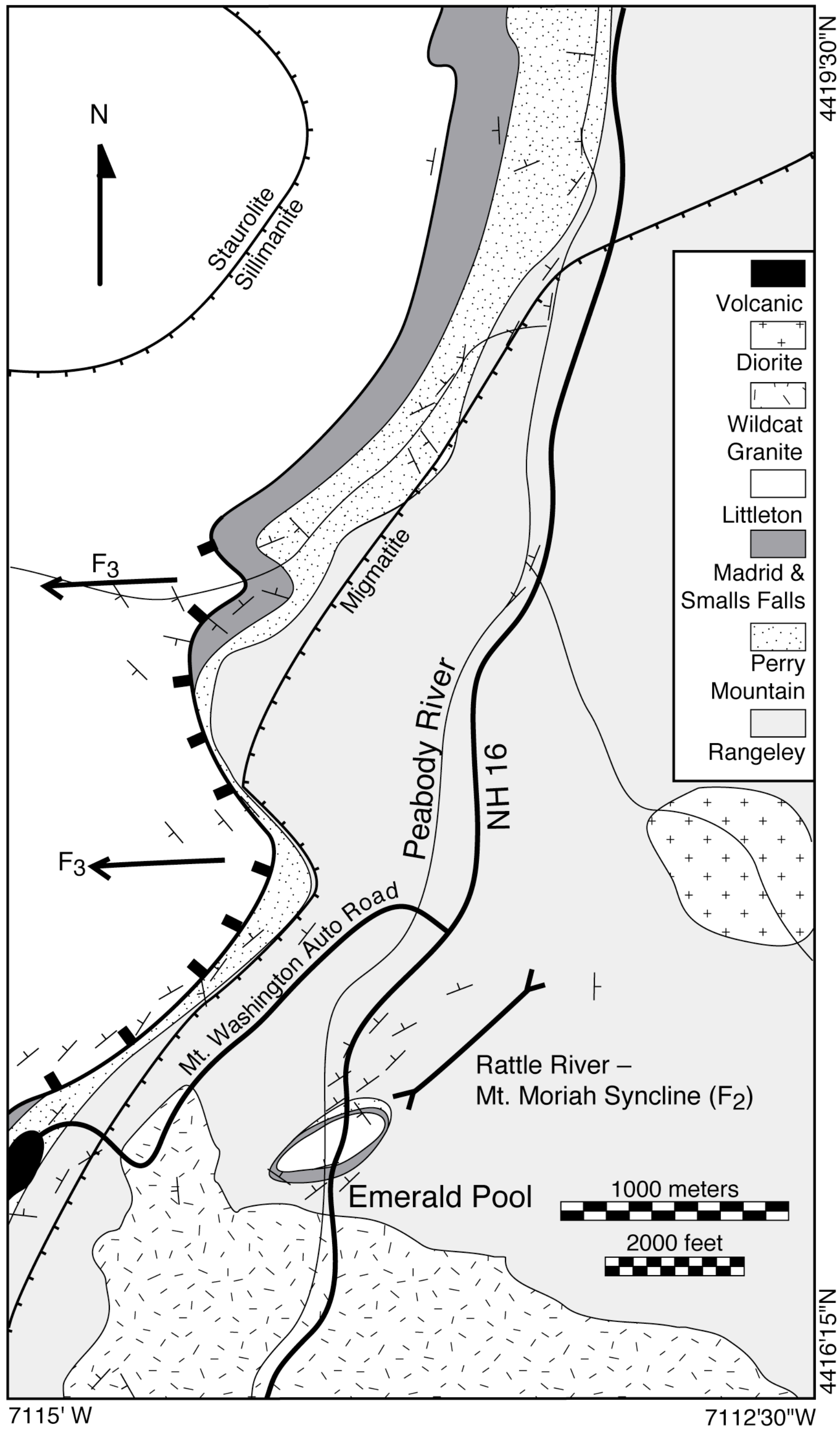

Figure 4: Geologic and structural map of the Pinkham Notch, NH, study area. F2 and F3 folds labelled. Rocks at Emerald Pool are part of a disrupted block of the Rattle River – Mt. Moriah Syncline. This area is the subject of Trip C6, this volume (see also Allen, 1996b).

distribution of poles to bedding and foliation planes in a equal area stereonet diagram, or pi diagram (Fig. 3A). This folding must be at least F2, as it folds a previous foliation. However, no F1 fold axes have yet been identified. This cryptic F1 folding event may have produced large recumbent isoclinal fold and thrust nappe structures, probably eastward vergent, similar to the F1 events observed elsewhere throughout New Hampshire (e.g., Eusden, 1988). F2 fold axes and lineations are plotted on an equal area stereonet in Fig. 3B, and plunge shallowly alternately to the northeast and to the southwest.

The F2 folds would have refolded the cryptic F1 nappes, and further mapping in the area may be able to identify these nappes through the map pattern. For example, the younger rocks (Littleton Formation) in the cores of these synclines do not cross the ridgelines or mountain tops. Where these synclines intersect ridgelines or mountain tops, older rocks (Rangeley Formation) outcrop. These older rocks form the upper plate of a recumbent synclinal nappe that has been refolded down into the cores of the crossing synclines.

The F2 folding event appears to have produced three major synclines, with intervening anticlines. Our traverse (Fig. 1) will encounter the first syncline in the upper reaches of the Rattle River north of Mt. Moriah, and the second syncline in the middle stretches of Moriah Brook. The third syncline is on the slopes between the Wild River and the Basin Rim. A small subsidiary syncline may occur between Howe Peak and Shelburne Moriah Mountain. These structures are difficult to trace along strike, some evidence in fact suggests that these structures may be discontinuous or disrupted.

This structural disruption is depicted in Fig. 1, notably by the block of rocks at Emerald Pool, shown as an extension of the F2 Rattle River—Mt. Moriah Syncline (Fig. 4). My interpretation differs significantly from previous maps of this area (Billings and others, 1941, 1975; Hatch & Wall, 1986). Past interpretations of these rocks have tried to relate them to the belt of Madrid and Smalls Falls just to the west (Fig. 4; the Boott Member of the Littleton Formation, Billings & Fowler-Billings, 1975), that separates the un-migmatized schists of the Littleton Formation on Mt. Washington from the migmatites of the Rangeley Formation in Pinkham Notch and to the east. Detailed mapping shows that the rocks at Emerald Pool are isolated within migmatites of the Rangeley Formation. There are numerous other examples of such isolated blocks along F2 fold trends (Fig. 1), with migmatites truncating the bedding of less-migmatized units. One possible explanation might be that of granitic magmas migrating upward through the crust preferentially along these upright F2 axial planes (Fig. 5; although pegmatite dikes in the area show no preferred orientation (Fig. 3C)). This upwelling may have driven the migmatization process (trip C6, this volume) creating highly mobile rocks; disrupting the F2 structures and skewing F2 fold axes (Fig. 3B).

Along the western boundary of the migmatite zone, structures are somewhat different, as there is a strong sense of folding about west directed fold axes (Fig. 4; and Billings & others, 1941, 1975). This folding is deemed to be a third generation of folding (F3), and probably post-dates the migmatization of the rocks to the east. The relative timing of the folding is indicated by the truncation of the migmatite front against map scale F3 folds (Fig. 4). Here also, the bedding and foliation planes dip only moderately to the west (Fig. 3D)—in the migmatite zone the dip tends to be much steeper (Fig. 3A). The poles to bedding and foliation planes from Pinkham Notch also roughly define a girdle on the equal area stereonet (Fig. 3D), that is consistent with the trend and plunge of fold axes and lineations in the area (Fig. 3E). A spaced cleavage, that may be axial planar to these F3 folds, is observed only along this western boundary of the migmatite zone. Poles to these cleavage planes are shown on an equal area stereonet projection in Fig. 3F. Comparison with Fig. 3D suggests that the cleavage is roughly parallel with the dominant trend of bedding and foliation.

There appears to be some fault motion, with cutting out of units, along the western boundary of the migmatite zone, perhaps associated with this F3 folding (Fig. 4). The axial planar cleavage of the F3 folds is roughly parallel to the presumed orientation of the fault surface. It is possible that this faulting and folding may represent downsliding of the un-migmatized rocks to the west related to the upwelling of granite magmas through the migmatite zone. On the other hand, the faulting may be related to intrusion of small necks of White Mountain Magma Series volcanic vent agglomerate that occur along this trend (Billings & others, 1975, 1979). Of course, these intrusions may have re-activated pre-existing structures related to the boundary of the migmatite zone.

DISCUSSION AND CONCLUSIONS

In summary, I recognize a four-stage structural history consisting of a possible east or northeast vergent isoclinal fold and thrust nappe (F1), refolded by a series of nearly upright anticlines and synclines with axes plunging gently alternately north-northeast and south-southwest (F2). These folds are disrupted and the F2 fold axes skewed apparently by intrusion of granitic magmas (probably along F2 axial planes) and the development of highly mobile partially melted migmatites. Finally, there is open folding about moderately westward plunging axes (F3) developing an axial planar spaced cleavage, along the western margin of the migmatite zone. Faulting along this boundary may be related to the F3 folding and the upwelling of granitic magmas in the migmatite zone, or may be due to later volcanic activity.

With the exception of late faulting, all the deformation and metamorphism in this area is presumed to be Acadian in age. Eusden & Lux (1994) report Ar40/Ar39 ages for metamorphic muscovites from this area (including migmatite outcrop #036 of Allen, 1992) of 300 to 275 Ma. The muscovite samples were collected over a vertical relief of 1.5 km, and suggest very slow cooling and uplift rates at that time (Eusden & Lux,1994). These results suggest that metamorphism and deformation in this area could not have been related to late stage magmatism, such as the Carboniferous Sebago Batholith (Osberg et al., 1985), but must be Acadian.

Eusden (1988) and Lyons et al. (1991) drew the trace of the Central New Hampshire Anticlinorium through the CM–WR area (Fig. 1). My mapping has confirmed that the metasedimentary rocks in this zone are predominantly the Silurian Rangeley Formation, the oldest unit in the Central Maine Terrain of New Hampshire. This is consistent with the placement of the Central New Hampshire Anticlinorium here. In addition, the dominant structures in the CM–WR zone are very nearly upright (Fig. 3A), which is consistent with the upright structures of Eusden's (1988) central "dorsal zone." Structures to either side of the central dorsal zone tend to be inclined or recumbent. Other migmatite zones similar to the one studied here, and other metamorphic "hot spots," are also centered on the Central New Hampshire Anticline or "dorsal zone" (Eusden, 1988; Chamberlain & Lyons, 1983; Chamberlain & Rumble, 1988). Most of the plutons of the Spaulding and Concord groups of the New Hampshire Plutonic Series are also found within the Central New Hampshire Anticlinorium. This suggests that anticlinorium structures forming the cores of orogenic belts may provide a structural control on pluton migration through the crust, or vice versa (Fig. 5). Additionally, this pluton migration might be responsible for the migmatization of these rocks (trip C6, this volume, Allen, 1992, 1996b). Comparison of the cartoon depicted in Fig. 5 with the map of the Acadian Orogen in Fig. 2, furthers the concept of the New England Appalachians as a surrogate crustal section (Chamberlain & Robinson, 1989; Rodgers, 1970, p. 114).

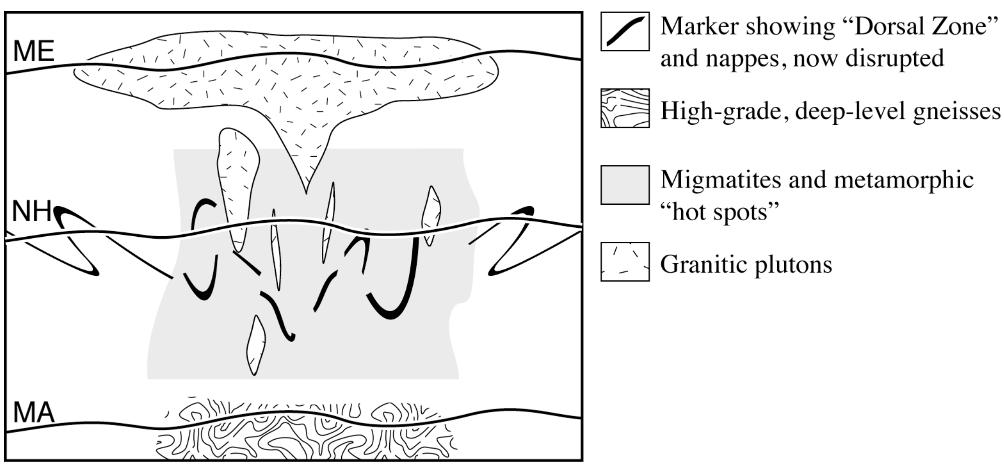

Figure 5: Cartoon depicting granitic magma migration from deep crustal levels through the crust along preferential pathways related to the "Dorsal Zone" (Eusden, 1988), disrupting structures and driving anomalous "hot spot" metamorphism and migmatization.

ACKNOWLEDGEMENTS

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"ROAD" LOG

MEETING POINT: Assemble (7:00 AM!) first at the Wild River Campground at the end of the Wild River Road off of the Evans Notch Road (ME/NH 113) (see (A) on Fig. 1), consolidate into as few vehicles as possible—leaving as many as possible at Wild River—and proceed to the Rattle River Trailhead (Appalachian Trail) on Route 2 three miles east of Gorham. Almost the entire trip will be on foot, over the Rattle River, Moriah Brook, and connecting trails (approximately 13 miles, with 3300 feet of vertical relief—"book time" of almost 9 hours). The area between the Carter-Moriah ridge and the Wild River is a federally designated Wilderness Area, with restrictions on the size of hiking groups. Be prepared to spend a full day hiking in the mountains—wear appropriate boots and clothing, and bing plenty of food, water, and extra clothing. At the end of the hike, we will consolidate into the vehicles we left at Wild River and return to Rattle River and thence to Bethel.

Maps (USGS 7.5 minute 1:24,000 quadrangles): Shelburne, NH-ME; Wild River, NH-ME; Carter Dome, NH. The AMC Carter-Moriah Range Trail Map and the AMC White Mountain Trail Guide are also useful. The best map for the Pinkham Notch area is Washburn's "Mount Washington and the Heart of the Presidential Range" (1988, 1:20,000).

Mount Moriah Traverse, Trail Mileage

0.0 Rattle River Trailhead (Appalachian Trail), three miles east of Gorham on US route 2. Proceed up the Rattle River trail parallel to the Rattle River (south).
1.7 Rattle River Shelter

STOP 1. PERRY MOUNTAIN or LITTLETON and RANGELEY (45 minutes): Outcrops in the river downstream of Shelter of pegmatite and aluminous gray migmatite gneiss with preserved bedding layers, possibly the Perry Mountain formation. At the shelter and upstream, extensive outcrops of orange Rangeley migmatite gneiss with pods, and extensive pegmatite intrusions.

3.2 Stream crossing below pool and falls

STOP 2. SMALLS FALLS, MADRID and/or LITTLETON (15 minutes): Bordering the pool, rusty and flaggy quartzites and schists of the Smalls Falls formation, and at the falls above the pool, layered gray aluminous schist cyclicly interbedded with fine laminations of granofels, possible graded bedding, of upper Madrid or lower Littleton affinity.

4.3 Rattle River trail ends, follow Kenduskeag trail right (west) towards Mount Moriah.
4.5 Trail slabs across the south side of Middle Moriah Mountain

STOP 3. MADRID and RANGELEY (15 minutes): small outcrops of light gray to grayish green tabular bedded calc-silicates, occasionally including dark green amphiboles. These rocks will persist as float as we move into gray to orange pod bearing migmatite gneiss with grit horizons.

5.7 Kenduskeag trail ends, follow short side trail to the summit of Mount Moriah (outcrops of Rangeley gneiss) then follow the Carter–Moriah Trail south towards Imp & the Carters.
6.7 Ledges overlooking the headwaters of Moriah Brook

STOP 4. RANGELEY (15 minutes): orange migmatitic gneiss, abundant pods generally aligned parallel to foliation, locally blood red (sulfurous) "pods" with a flaggy schistose foliation, abundant diffuse pegmatites, and possible zones of re-crystallized quartz pebble conglomerate.

7.1 Stony Brook trail comes in from right (west). Follow the Moriah Brook trail left (east) towards the Wild River.
9.8 Trail crosses from northeast bank to southwest bank of Moriah Brook

STOP 5. LITTLETON and MADRID within RANGELEY (45 minutes): upstream from where the trail crosses the stream, rhythmically bedded aluminous schist and quartz granofels and laminated green calc-silicates, cut at a low angle by migmatite gneiss.

11.2 Trail crosses from southwest back to northeast bank of Moriah Brook

STOP 6. LITTLETON CONGLOMERATES, MADRID, and SMALLS FALLS above RANGELEY (45 minutes): gray pebble/cobble conglomerates, gray to rusty brown schists, and laminated green-brown calc-silicates with 0.1 to 1 meter-scale chevron and similar folding. Within the gorge are punky weathering rusty sulfidic schists, and pod bearing orange migmatite gneiss, as well as abundant pegmatites.

12.2 Highwater Trail joins from the right (southwest), follow the Moriah Brook trail northeast, to the Wild River.
12.4 Outcrop in Wild River of orange to gray migmatite gneiss with pods of sulfidic rusty schist and possible relict bedding represented by thin granular horizons separated by aluminous horizons.
12.6 Cross bridge over the Wild River. Follow the Wild River Trail northeast 0.3 miles to campground and autos.

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