THE NOTCHES:
BEDROCK AND SURFICIAL GEOLOGY OF NEW HAMPSHIRE'S WHITE MOUNTAINS

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This trip will travel through the spectacular scenery of Pinkham, Crawford, and Franconia Notches in the White Mountains of New Hampshire. These ranges include the highest peaks in the northeastern United States as well as many famous landforms such as The Old Man of the Mountain and Tuckerman Ravine. We will examine the entire geologic history exposed here, from the Silurian to the Holocene, including the nature of sedimentation, deformation, metamorphism, and magmatism during the Devonian Acadian orogeny; the Mesozoic magmatic episodes associated with hot-spot migration and rift-drift crustal extension during the opening of the Atlantic; and the Quaternary history of alpine and continental glaciation and deglaciation, as well as fluvial and landslide activity.

SILURO-DEVONIAN BEDROCK GEOLOGY OF THE PRESIDENTIAL RANGE

The Presidential Range of northern New Hampshire lies within the White Mountain National Forest and is one of the most popular hiking areas in the country. Culminating with Mount Washington (elev. 6288 ft.), the range contains the most extensive alpine zone in the eastern U.S. as well as rugged and remote wilderness areas. Eusden and students from Bates College (1996a,b) have been mapping the bedrock geology of the Presidential Range in an effort to better understand the effects of the Devonian Acadian orogeny. Allen (1992, 1996a,b) has mapped the bedrock geology in Pinkham Notch and the Carter-Moriah Range to the east. Our goals are to investigate the variations in stratigraphy, deformation sequences, and metamorphism that occur in the Silurian and Devonian metasedimentary cover rocks of the Central Maine Terrane.

Geologic setting

The Presidential Range is located on the western flank of the Central Maine Terrane (Fig. 1), the Silurian and Devonian cover rocks of which correlate with both the Central Maine Basin of Bradley et al. (1998) and the Merrimack Belt of Robinson et al. (1998). The Silurian and Devonian cover rocks of the Central Maine Terrane stretch from Connecticut to New Brunswick and are bounded to the southeast by composite Avalonian rocks along the Maine, New Hampshire, and Massachusetts coasts and to the northwest by the Bronson Hill, Boundary Mountains, and Lobster Mountain anticlinoria (Lyons et al., 1997, Bradley et al., 1998, and Robinson et al., 1998).

The Central Maine Terrane contains Silurian metasedimentary cover rocks which are interpreted as an eastward thickening sequence of deep water turbidites deposited in either a passive margin basin (Robinson et al., 1998; Moench & Pankiwskyj, 1988) or a forearc basin associated with a northwest dipping subduction complex (Hanson & Bradley, 1989; Bradley et al., 1998; Eusden et al., 2000). The Silurian rocks in contact with and adjacent to the Bronson Hill and Boundary Mountains anticlinoria are thin, near-shore conglomerates and calcareous turbidites of the Clough, Fitch, and portions of the Rangeley Formations. These thicken to the southeast into deeper water turbidites of the Rangeley, Perry Mountain, Smalls Falls, and Madrid Formations (Hatch et al., 1983; Moench & Pankiwskyj, 1988; Hanson & Bradley, 1989, 1993). Conformably overlying these rocks are Devonian deep water turbidites of the Littleton, Carrabassett and Seboomook Formations. These formations were deposited in a foreland basin setting associated with either a southeast dipping subduction system that overrode the Silurian northwest dipping subduction system (Bradley et al., 1998) or a double subduction that consisted of the same northwest-dipping Silurian subduction system that persisted into the Devonian and the southeast system of Bradley et al. (1998) (Eusden et al., 2000).

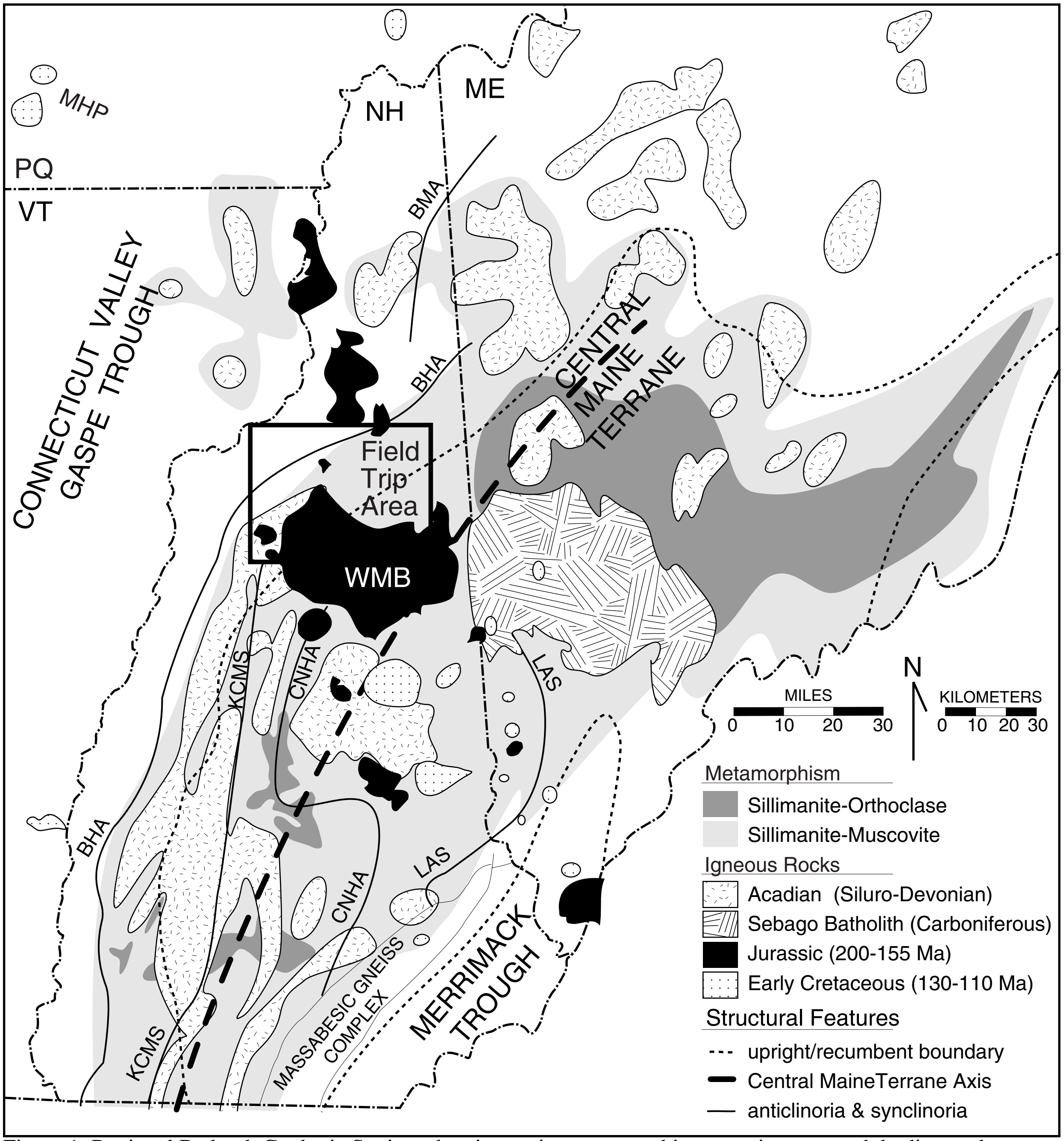

Figure 1: Regional Bedrock Geologic Setting, showing major metamorphic zones, igneous rock bodies, and structural features, and the field trip area. WMB = White Mountain Batholith. MHP = Monteregian Hills Province. BMA = Boundary Mountains Anticlinorium. BHA = Bronson Hill Anticlinorium. KCMS = Kearsarge-Central Maine Synclinorium. CNHA = Central New Hampshire Anticlinorium. LAS = Lebanon Antiformal Synclinorium.

The Central Maine Terrane has experienced intense ductile deformation, high-grade metamorphism, and a protracted period of pre- syn- and post-kinematic granitic plutonism. In general, the deformation in the northeast part of the Central Maine Terrane (Maine and New Brunswick) is dominated by upright structures and lower grade post-kinematic contact metamorphisms associated with syn-, but largely post-kinematic plutons (Moench & Pankiwskyj, 1988; Osberg et al., 1989; Guidotti, 1989). In the regions to the southwest (portions of western Maine, New Hampshire, Massachusetts, and Connecticut) structures are generally recumbent, multiply deformed, and accompanied by syn-kinematic higher-grade metamorphisms and associated intrusion of predominately syn-kinematic granitic plutons (Lyons et al., 1997; Eusden & Lyons, 1993; Robinson et al., 1998). The transition between these different styles of the Acadian tectonism occurs in a zone only 75-100 km long, as measured along the strike of the Central Maine Terrane, and are attributed to a transition from shallower crustal levels to deeper crustal levels through the orogen that are now exposed (Carmichael, 1978; Osberg et al., 1989). The Presidential Range straddles this fundamental transition in the Acadian orogen. Modeling of granite ascent in convergent orogenic belts by Solar et al. (1998) and Solar & Brown (1999), and new geochronology on syntectonic intrusive rocks by Bradley et al. (1998), suggests that the synchronous nature of deformation, metamorphism, and plutonism, and the zone of the transition, may extend into western Maine.

Bedrock geology of the Presidential Range

Using as a foundation the excellent work of previous geologists who have mapped the rocks in the Presidential Range (Billings, 1941; Billings & Fowler-Billings, 1975; Billings et al., 1946, 1979; Hatch & Moench, 1984; and Moench et al. 1999), we are nearing the end of ongoing mapping project that has taken 11 years to date, to redefine the geology in the range. Figure 2 shows the geologic map we have compiled.

We have subdivided the Devonian Littleton Formation into 16 members, recognize both the Silurian Madrid and Smalls Falls Formations, subdivide the Rangeley Formation into 9 members, and recognize an episode of sedimentary disruption in the Rangeley which we interpret as a migmatized olistostromal melange (Eusden et al, 1996a).

The sequence of deformation is interpreted to comprise six events numbered D0 through D5. D0 is a phase of pre-metamorphic faulting. D1 is characterized by east verging isoclinal nappes. D2 is characterized by the Clay klippe and Greenough Spring thrust (see Eusden et al., 1996a). D3 folds result in anomalous easterly dips of bedding (S0) and D1 foliation (S1), macroscopic refolding of the D2 thrusts, and define the great Chandler Ridge Dome. D4 folds are the most common structural features seen in the Presidential Range and vary in scale from mesoscopic at high elevations to meso- and microscopic at lower elevations. D5 is principally a crenulation that is restricted to the Pinkham Notch region. The many vertical and lateral variations in structural style that the phases of deformation exhibit further reflect the complex nature of the deformation that occurred in the Acadian crustal transition (Eusden et al., 1996a).

Many pulses of metamorphism have been recognized in the Presidential Range. The following summary is based on work of Eusden et al. (1996b), Wing (1996), Hatch & Wall (1986), and Allen (1992, 1996a,b). M1 is characterized by aligned andalusite, much of which is now preserved as pseudomorphs, and occurred during D1 nappe-stage folding. M2, the peak of metamorphism at circa 404 Ma. (Eusden et al., 2000), is characterized by sillimanite zone metamorphism in the Littleton schists and migmatization in the Rangeley gneisses and occurred during the later part of D1 nappe-stage folding. The later events, M3 and M4, are contact metamorphic events both reaching staurolite grade. M3 occurred prior to D4 folding based on the observation that granites related to this phase of metamorphism are folded by F4 folds. M4 metamorphism occurred after D4 deformation and is related to the latest stage of post-tectonic granite intrusion (circa 355 Ma.) and D5 crenulation. M5 is a retrograde metamorphism producing scattered occurrences of chlorite and/or sericite alteration in the schists and gneisses.

The earliest plutonism in the Presidential Range is characterized by one rare dioritic stock, the circa 408 Ma. Wamsutta Diorite (Guzofski, 1997; Eusden et al, 2000). This pluton has a weak S1 foliation but also cuts across the S1 fabric in the metasedimentary rocks. The diorite represents the earliest syn-kinematic intrusion in the range, and is probably in part synchronous with M1. A widespread scattering of sills, veins, and small plutons of two-mica granite intruded subsequently. M2 and M3 metamorphisms were associated with the intrusion of these granites. These granites are deformed by F4 folds. Allen (1992, 1996a,b) mapped the largest of these granites, the circa 401 Ma. (Bowring, pers. com. cf. Anton, 1998) Wildcat Granite, near the Wildcat Ski Area immediately adjacent to the Presidential Range. Late stage post-tectonic granites are restricted to the Peabody River Stock previously identified by Allen (1992, 1996a,b), Hatch & Wall (1986), and Billings & Fowler-Billings (1975), and recently dated at circa 355 Ma by Eusden et al. (2000).

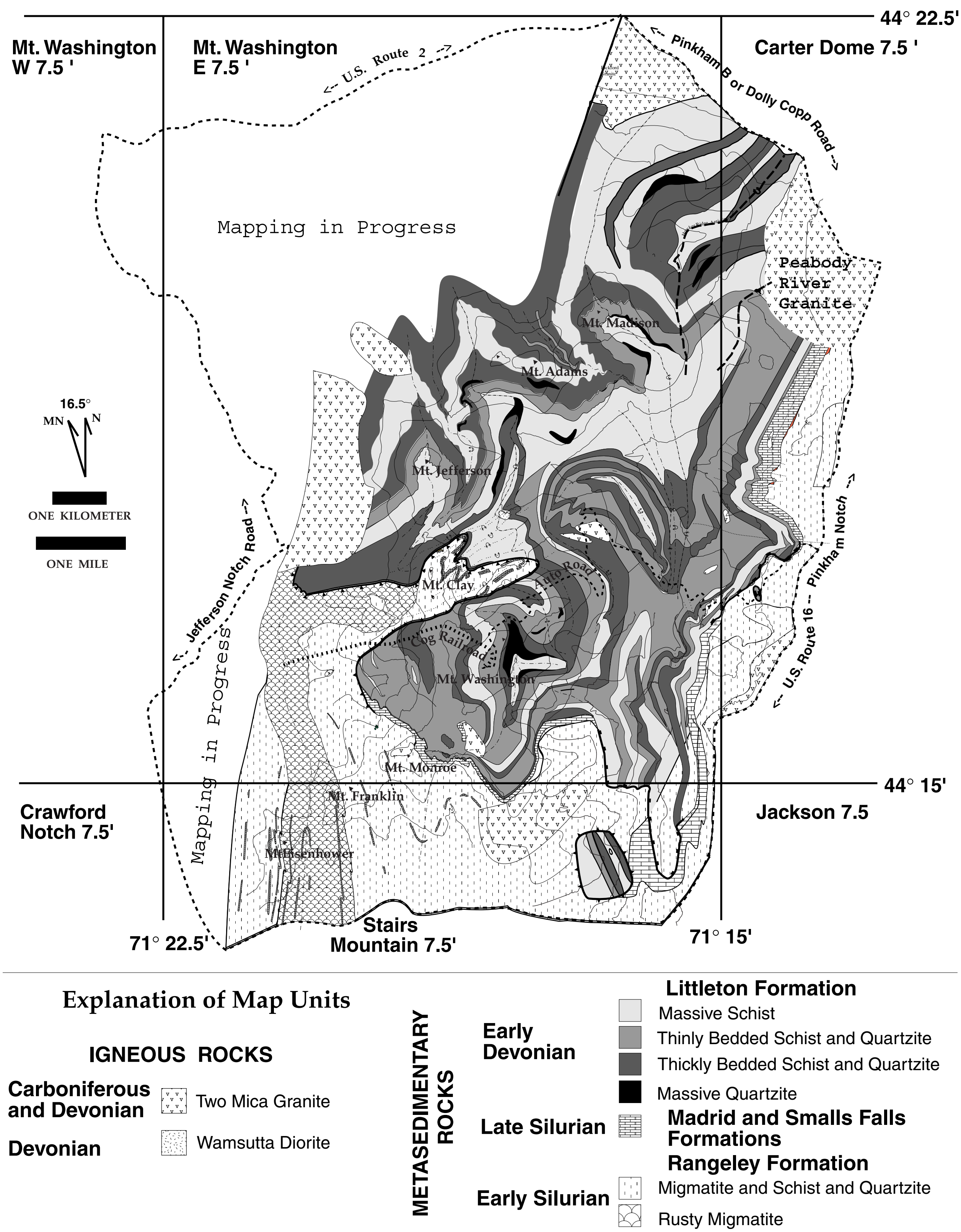

Figure 2. Generalized bedrock geology in the Presidential Range, New Hampshire.

U-Pb geochronology from the Presidential Range constrains the time of several key events in the history of the Acadian orogeny of northern New Hampshire (Eusden et al., 2000): 1) the boundary between the deforming orogenic wedge and the depositional site in the foreland basin was located in the vicinity of the Presidential Range at 408 Ma.; 2) southeast verging thrust-nappes formed at circa 408 Ma. and plate tectonic models for the Acadian must be modified to account for both a northwest-migrating deformation front and a southeasterly verging thrust-nappe belt; 3) peak metamorphism in schists, migmatization, and intrusion of early two-mica granites occurred at circa 404 Ma.; 4) the younger limit of Paleozoic orogenesis in the Presidential Range is at 355 Ma as recorded by the post-kinematic Peabody River Granite. From circa 408 to 404 Ma., and perhaps as late as circa 394 Ma., the Acadian orogeny in the Presidential Range was typical of the syn-kinematic style seen in central and southern New Hampshire, Massachusetts, and Connecticut. From no earlier than circa 394 Ma to as late as circa 355 Ma, the tectonism was typical of the post-kinematic style seen in parts of Maine. All of these constraints confirm the presence of a complex system of temporal and spatial variations in the styles of tectonometamorphism that occurred within the Acadian transition zone of the Northern Appalachians.

MESOZOIC MAGMATISM OF THE WHITE MOUNTAIN IGNEOUS PROVINCE

The White Mountain igneous province (magma series of Billings, 1934) consists of plutons, ring complexes, and volcanics emplaced along a NNE trend across New England (Fig. 1). Four petrographic associations are recognized (Creasy, 1974; Eby, 1987): (1) alkali syenite-quartz syenite-granite; (2) subaluminous biotite granite; (3) gabbro-diorite-monzonite and; (4) syenite-nepheline syenite. The igneous activity is largely confined to two periods, 200-165 Ma and 130-110 Ma (Eby & others, 1992). McHone & Butler (1984) relate these two major periods of igneous activity to the opening of the North Atlantic Ocean. The reader is referred to Eby (1987) for an overview of the White Mountain igneous province. The older White Mountain igneous province is dominated by silica-oversaturated subaluminous to peralkaline rocks of association 1, including the White Mountain batholith. Two minor nepheline-bearing intrusions occur at Red Hill, New Hampshire, and Rattlesnake Mountain, Maine. These two occurrences together with nepheline-bearing intrusions of the younger White Mountain province define a narrow zone that strikes at high angle to the NNE trend of the overall province (Fig. 1: Creasy, 1989). To the north of this zone are found the large composite plutons and batholith of the older province; to the south only a few small, scattered plutons of this age are present. In contrast, nearly all plutons of the younger White Mountain province are found to the south of this zone.

The Monteregian Hills and younger White Mountain igneous provinces represent the last period of igneous activity in New England (130-100 Ma). The bulk of the magmatism occurred Ca. 125 Ma, but younger ages have been obtained for Little Rattlesnake (114 Ma, Foland & Faul, 1977) and Cuttingsville (100 Ma, Armstrong & Stump, 1971). Plutons emplaced to the west of Logan's line consist largely of mafic alkaline suites, many of which are nepheline normative. To the east of Logan's line, felsic rocks are much more important components of the intrusions and silica-undersaturated rocks are not found. Some of these younger plutons show ring-like structures (Ossipee and Pawtuckaway) while others appear to be small plugs (e.g. Little Rattlesnake, Ascutney and Tripyramid). In most cases the most evolved rocks are syenites and quartz syenites, but biotite granites are found at Ossipee and Merrymeeting Lake.

On this field trip, we will focus on components of the White Mountain batholith [associations 1 and 2 above] which comprises about 50% of the total areal extent of the older White Mountain igneous province. The material in this field guide previously appeared in Creasy & Eby (1993) and Creasy & Fitzgerald (1996). Consult either of these sources for further information.

The White Mountain Batholith

The White Mountain batholith (Fig. 3; see also Hatch & Moench, 1984) is a composite of several overlapping centers of felsic magmatism. Individual centers are strikingly defined by composite ring dikes of porphyritic quartz syenite. Thick Sections of rhyolitic crystal tuffs, breccias, and subvolcanic granite porphyry are partially circumscribed by the ring dikes. A mosaic of subalkaline to peralkaline silica-oversaturated plutons intrude these centers and provide areal continuity to the batholith. Distribution of porphyritic, miarolitic, and aplitic textures indicate that the roofs of several plutons are partially intact.

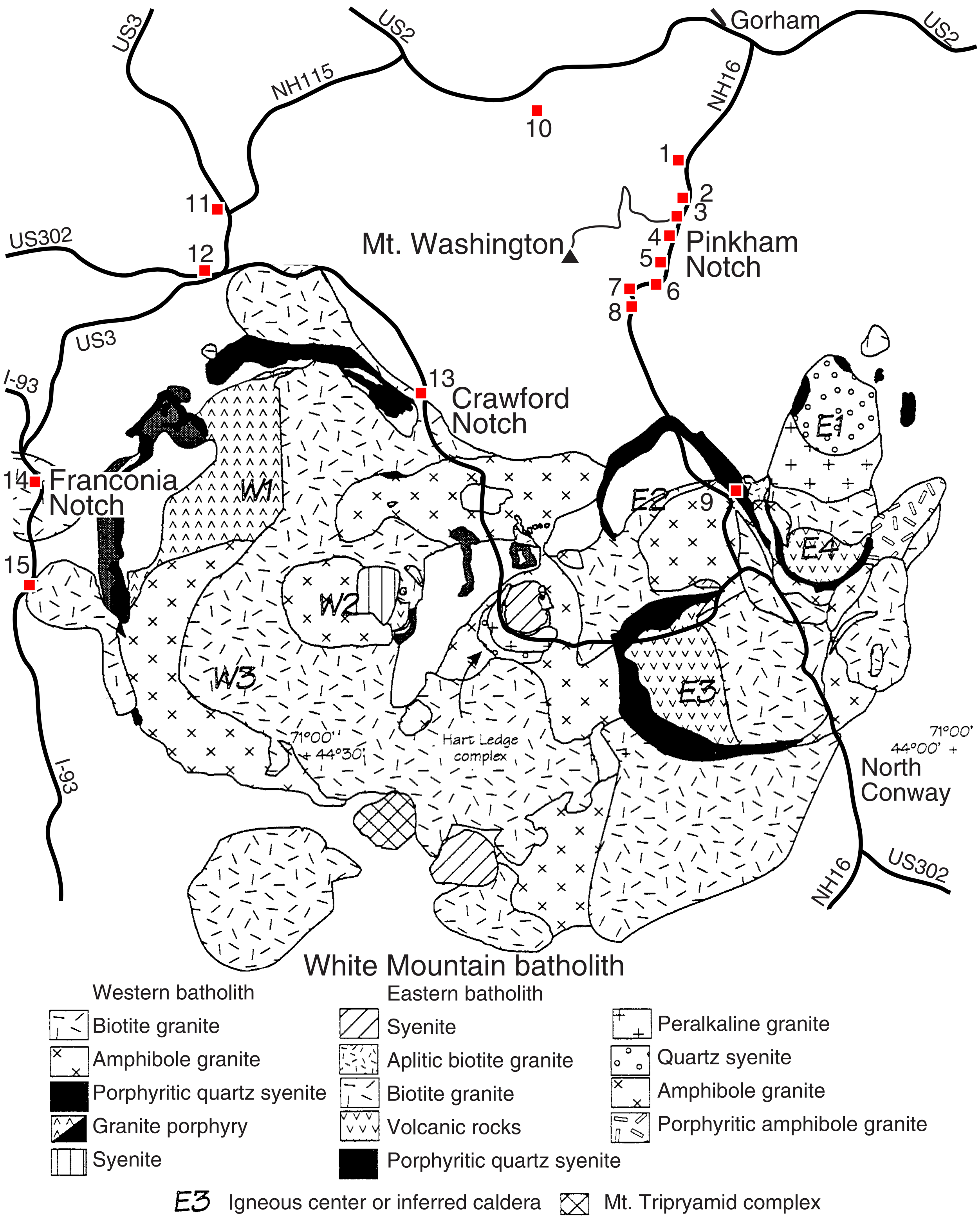

Figure 3: Geologic map of the White Mountain batholith (after Creasy, 1974; Osberg et al., 1978), with locations of field trip stops (numbered squares) and main highways.

The geology of the White Mountain batholith is described by Billings (1928), Billings & Williams (1935), Creasy (1974), Davie (1975), Eby & others (1992), Fitzgerald (1987), Henderson & others (1977), Moke (1946), Osberg &

others (1978), Parnell (1975), Smith & others (1939), Wilson (1969), and Wood (1975). Granites, quartz- syenites, and syenites account for about 97% of the 1,000 km^2 area of the batholith; volcanic rocks of similar composition account for the remainder. Pink, medium-grained subalkaline biotite granite (the Conway Granite) and a green, medium-grained subalkaline to peralkaline amphibole granite (the Mount Osceola Granite) comprise 80% of the batholith. Medium-grained sub-alkaline to peralkaline amphibole syenites and quartz syenites are widely distributed and are similar in occurrence, texture, and mineralogy to the Mt. Osceola. Distinctive porphyritic quartz syenite occurs in ring dikes in the western (the Mount Garfield) and the eastern (the Albany) halves of the batholith.

Volcanic rocks (the Moat Volcanics), chiefly trachyte, tuff, breccia, and alkali rhyolite and comendite, are found in the eastern portion of the batholith. Only minor occurrences of such lithologies are present within the ring dikes of the western batholith. Several units of granite porphyry (grouped as the Mount Lafayette unit) occurring in the western batholith differ little in texture or mineralogy from the comendites of the eastern batholith. We show these rocks as volcanics (Fig. 3) although definitive volcanic textures are generally lacking. Historically, the granite porphyry is treated as intrusive and not included within the Moat Volcanics (Billings & Williams, 1935; Eby & others, 1992).

Gabbro, diorite, and monzonite are present in the Mount Tripyramid complex (Fig. 3), a member of the younger White Mountain igneous province, that is spatially associated with the White Mountain batholith.

Emplacement of the White Mountain Batholith

Emplacement of the batholith occurred in middle and late Jurassic time, 201-155 Ma (Eby & others, 1992). The western half of the White Mountain batholith, exposed in the Franconia and Crawford Notch 15’ quadrangles, contains three igneous centers (W1-3, Fig. 3), the largest of which is 20 km in diameter. Igneous activity commenced in the western batholith with emplacement of the porphyritic quartz syenite (201 Ma and 193 Ma) and the quartz porphyry (195 Ma) of center W1 and the syenite and trachyte of center W2 (193 Ma). Subsequent intrusions of amphibole granite (187 Ma) [possible W3 (?)] and biotite granite (181 Ma) were widespread across the entire area of the batholith. Intrusion of peralkaline granite (177 Ma) in the eastern part of the pluton is considered an extension of the amphibole granite (Mount Osceola) event.

The eastern portion of the White Mountain batholith, exposed in the North Conway and Crawford Notch 15’ quadrangles, has at least four magmatic centers (Fig. 3). Two centers with thick pyroclastic successions are interpreted as calderas (Noble & Billings, 1967; Fitzgerald, 1987; Fitzgerald & Creasy, 1988). Other centers where ring dikes or crescent-shaped intrusions are associated with epizonal plutons define more deeply eroded calderas. Caldera development here post-dates similar events in the western batholith by about 10-20 Ma. Dated units include the ring dike of center E2 (179 Ma); the Moat Mountain volcanic sequence (173-168 Ma) and ring dike (170 Ma) of center E3; and plutons of biotite granite (171 Ma and 155 Ma). We interpret the White Mountain batholith as a subhorizontal slice through a caldera field cut about 1.5 km thick and 1-2 km below the original land surface.

GLACIAL GEOLOGY OF THE NORTHERN WHITE MOUNTAINS

Introduction and previous work

The glacial history of the White Mountains has been the subject of much controversy. During this field trip we will consider new evidence that has forced a revision of long-standing. The following history of glacial studies in the White Mountains is summarized from W.B. Thompson (1999). Several of the most recent investigations of Quaternary geology in this part of New Hampshire were published in a special issue of *Géographie physique et Quaternaire* (v. 53, no. 1, 1999).

Soon after Louis Agassiz came to the United States in 1846, he traveled to the mountains with a group of Harvard University students to look for evidence of glaciation (Lurie, 1988). Agassiz found moraines and other features that indicated the region had been covered by glacial ice (Agassiz, 1870). Alpheus Packard, Jr. (1867a,b) and George Vose (1868) proposed that a central ice cap in the White Mountains fed a radiating series of valley glaciers. Vose noted the NE-SW trending striations in the Peabody River valley between Pinkham Notch and Gorham, but interpreted them as supporting the downvalley flow of a glacier originating in the Presidential Range. Charles Hitchcock (1878) disproved this theory by finding stoss-and-lee bedrock topography showing that continental ice flowed southwestward up the Peabody River valley. Some of the outcrops examined by these workers can still be seen along NH 16 near the entrance to the Mt. Washington Auto Road and will be noted during this trip.

Hitchcock (1876) also found evidence that the summit of Mt. Washington had been glaciated. Both Hitchcock and Agassiz (1870) realized that a continental ice sheet had advanced over the White Mountains from the north, but they thought that an ice cap persisted in the mountains after recession of the ice sheet. James W. Goldthwait (1916) refuted this theory by presenting evidence, including the provenance of erratic boulders and the distribution of moraines and glacial-lake deposits, showing that continental ice flowing from the north was the dominant type of glaciation.

Most early investigators assumed that dynamic ice persisted in the White Mountains during deglaciation. Goldthwait (1925) initially agreed that the northward-receding ice margin remained active and deposited moraines. However, in the 1930’s there was a fierce debate over the relative importance of active vs. stagnant ice during glacial recession from the White Mountains. Richard Flint’s (1930) work in Connecticut was a major influence that swayed most people toward the stagnation camp (R. P. Goldthwait, 1939a). Goldthwait (1938) eventually abandoned the concept of active-ice retreat in New Hampshire and discredited many of the moraines and other features described in his 1925 volume.

The deglaciation controversy peaked around 1940. In the face of growing opposition, Richard Lougee (1940) continued to defend the active-ice deglaciation model. Ernst Antevs (1939) and Douglas Johnson (1941) tried to show that the “normal retreat” and “downwasting” models were overly simplistic in their portrayal of a receding ice margin in high-relief terrain. Nevertheless, the concept of wholesale ice stagnation in New Hampshire prevailed through the middle 1900’s (Goldthwait et al., 1951) and has only recently been reevaluated in the White Mountains.

Recent studies in the northern White Mountains

Detailed surficial geologic mapping in New England has shown that the late Wisconsinan ice margin receded systematically, and that live ice persisted over much of the region during deglaciation (Koteff & Pessl, 1981). These findings contradict the earlier stagnation model. End moraines and other features support the presence of late-glacial active ice in northern New Hampshire, especially where valleys were favorably oriented for sustaining the flow of the thinning ice sheet.

In the northeastern White Mountains, Thompson & Fowler (1989) documented the Androscoggin Moraine east of Gorham. This moraine complex includes numerous sharp-crested bouldery ridges up to 30 m high, deposited by a tongue of the Laurentide Ice Sheet extending down the Androscoggin River valley to the Maine border. Gerath et al. (1985) likewise found evidence of late-glacial ice activity in the Success Moraine on the proximal side of the Mahoosuc Range northeast of Gorham.

Part of this trip will examine the Ammonoosuc and Israel River basins in the northwestern part of the mountains, where end moraines and other deposits associated with ice-dammed glacial lakes support the active-ice deglaciation model (Thompson et al., 1999). New work by Ridge et al. (1999) has updated the varve record in the nearby upper Connecticut River valley and its relationship to the deglaciation history of surrounding areas. Recent analysis of the vegetation history and radiocarbon dating of lake sediments in the White Mountains have been carried out by other investigators (Miller & Thompson, 1979; Davis et al., 1980; Spear, 1989; Thompson et al., 1996; Miller & Spear, 1999; Cwynar & Spear, 2001). These studies provide important data for interpreting the chronologic and climatic context of moraine systems and the sequence of ice recession.

The White Mountain Notches

Pinkham, Crawford, and Franconia Notches, from east to west (Fig. 3), are the largest examples of glacial troughs in the White Mountains, and all three are accessible from highways. Other smaller but no less spectacular notches in the

White Mountains include, from east to west, Grafton, Evans, Mahoosuc, Carter, Carrigain, Zealand, Hancock, and Kinsman Notches, some of which are only accessible by hiking trails.

Each of these notches was a mountain pass prior to multiple episodes of continental glaciation during the Pleistocene Epoch. The three notches that we will visit on this trip are substantially deeper than other higher elevation notches in the White Mountains because glacial ice moved through them for a longer period of time and/or was more erosive during each glacial episode.

Northwest of the White Mountain highlands, continental ice had few topographic obstacles interfering with its southeastward advance until it got to the steep northwesterly slopes of the White Mountains. Because continental ice sought the paths of least resistance through the mountains, it converged first in the deeper pre-existing passes through the mountains. As continental ice thickened during each episode, it began to move through the higher-elevation passes and eventually covered all the peaks at least once. During glacial maxima, converging ice flow (ice streams) within the continental icesheet probably existed within the notches.

The Bethlehem Moraine complex

As originally mapped by J. W. Goldthwait (1916), the Bethlehem Moraine consists of dozens of hummocks and ridges in the Ammonoosuc River basin from Littleton east to the Wing Road area in Bethlehem (just west of the field trip area). The moraine ridges range in height from 3 m to over 30 m, and are up to 1300 m long. They commonly trend between east and northeast. There are few bedrock exposures within the Bethlehem Moraine complex, but striated outcrops immediately to the north and south indicate ice flow directions of 175-185° (Goldthwait, 1916; W. B. Thompson, unpublished data).

Surficial deposits in the moraine complex are generally thick (up to at least 50 m). The moraine ridges are composed of silty-sandy till that locally contains lenses of sand, gravel, and silt. Large boulders of granite and gneissic granite are very abundant on the surfaces of many moraine segments. The elevations of the ridges suggest they were deposited in glacial Lake Ammonoosuc (described below), and this interpretation is supported by occurrences of deformed glaciolacustrine sediments within the moraine complex.

The morainic deposits of the Bethlehem area have been the focus of major controversies involving the modes of glaciation and ice retreat in the White Mountains. Early workers accepted them as moraines deposited by active ice, but ice-stagnation proponents of the middle 1900's opposed this concept. Thompson et al. (1999) reexamined the Bethlehem Moraine complex and correlative deposits to the east. They concluded that the numerous subparallel till ridges are indeed end moraines. Together with the associated deposits and spillways of glacial Lake Ammonoosuc, the moraines indicate oscillatory northward retreat of a coherent ice margin during late Wisconsinan deglaciation. The collective stillstands and minor ice readvances that formed these moraines are known as the Littleton-Bethlehem Readvance.

Glacial Lake Ammonoosuc

Glacial lake deposits are among the most significant deglaciation features in the northern White Mountains. Ice-dammed lakes existed in north and west-draining valleys in this area, and most of the Bethlehem Moraine was deposited in a lake that occupied the Ammonoosuc River valley. This water body was named "Lake Ammonoosuc" by Goldthwait (1916). It resulted from damming of the upper part of the valley when the late Wisconsinan ice stood in the Littleton-Bethlehem area. As the ice margin receded northward, successively lower spillways for the lake were uncovered and the lake level fell. The first stage of the lake (Crawford Stage) drained eastward through Crawford Notch, with a spillway at an elevation of about 573 m (1880 ft). This was the same route taken by an earlier subglacial drainage that formed the esker in the upper Ammonoosuc Valley (Goldthwait & Mickelson, 1982).

Lougee (n.d.) and Thompson et al. (1999) described the later spillways for Lake Ammonoosuc. Following the Crawford Stage, the lake drained southwestward through five progressively lower spillways into the Gale River valley in the Franconia area (Fig. 4). The spillway for the Gale River 2 Stage is a prominent channel that will be seen during this trip along US 3 southwest of Twin Mountain. This is the lake stage into which the large ice-contact delta

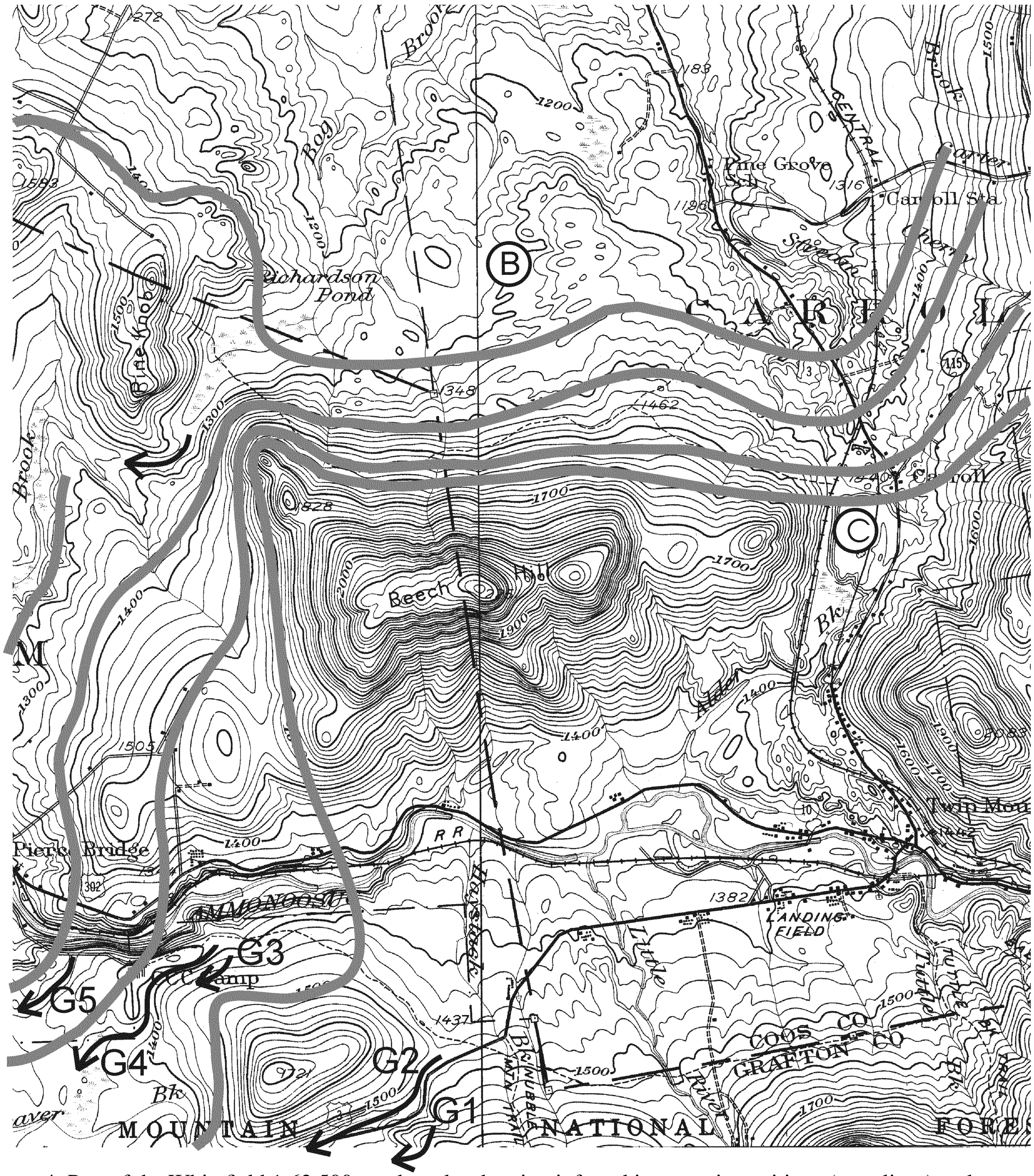

Figure 4: Part of the Whitefield 1:62.500 quadrangle, showing inferred ice-margin positions (gray lines) and meltwater spillway channels (arrows). G1-G5 are spillways for successive Gale River stages of glacial Lake Ammonoosuc. B: Beech Hill moraines. C: Carroll delta. From Thompson et al. (1999).

at Carroll was built (Stop 10). Later spillways north of Bethlehem village drained the Bethlehem and Wing Road Stages of Lake Ammonoosuc into Indian Brook and finally into the lower Ammonoosuc River.

During the evolution of glacial Lake Ammonoosuc, water entered the lake not only from the melting ice sheet, but also from the early postglacial Ammonoosuc River and smaller streams draining the surrounding mountains as shown by Lougee in Fig. 5. Much of the sediment deposited in Lake Ammonoosuc probably came from these non-glacial sources. The mouth of the river shifted farther westward as the lake level dropped, and the lake ultimately disappeared when the ice margin receded from the Bethlehem area.

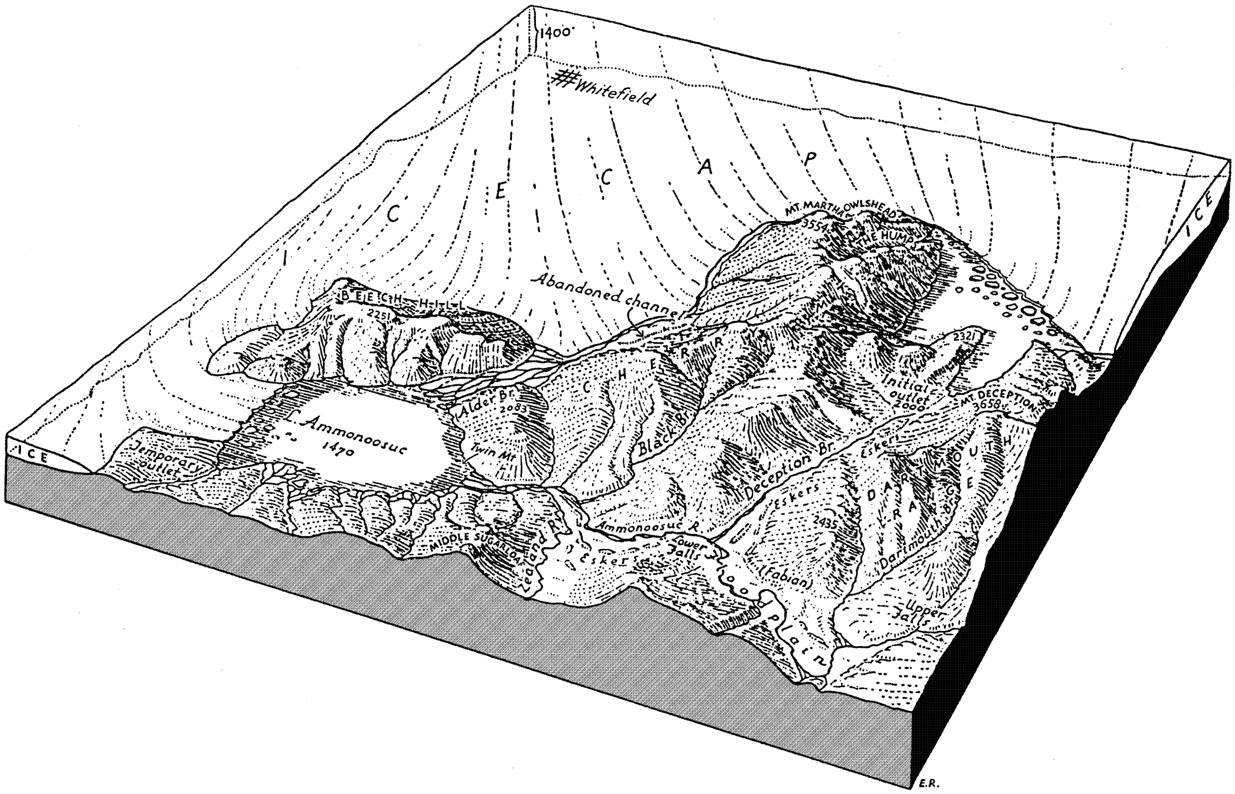

Figure 5: Diagram from Lougee (1940), showing location of ice margin when the Carroll delta was deposited into Gale River 2 stage of glacial Lake Ammonoosuc.

Deglaciation chronology

Regional chronologies of deglaciation for northern New England and adjacent Canada have been published by Davis & Jacobson (1985), Dyke & Prest (1987), and Ochietti (1989). These authors synthesized radiocarbon ages providing minimum limits on the time of ice retreat, and they prepared paleogeographic maps showing the inferred extent of glacial ice at various times following the late Wisconsinan glacial maximum. Their reconstructions show northern New Hampshire being deglaciated between 14,000 and 13,000 14C yr BP, followed by recession of the Laurentide Ice Sheet into southeastern Quebec, and then by marine transgression in the St. Lawrence Lowland (Quebec City area) shortly after 12,000 yr BP. The New England chronology is based on ages from terrestrial and a few marine sites (Davis & Jacobson, 1985), while the invasion of the Champlain Sea in adjacent Quebec is constrained by radiocarbon ages on marine shells from near the upper limit of submergence (Parent & Occhietti, 1988).

Figure 6 shows the locations of ponds or marshes from which sediment cores have been recovered to obtain radiocarbon ages summarized in Table 1. Only the oldest ages from sediment cores that most closely approximate the time of deglaciation at each site are presented. The decimal numbers on Fig. 6 are ages in thousands of radiocarbon years BP, and numbers in parentheses correspond to the site locations indicated in Table 1.

Radiocarbon ages from the basal portions of pond sediment cores yield minimum estimates for the time of deglaciation in northern New Hampshire (Fig. 6; Table 1). Pond of Safety in Randolph, just north of the Presidential Range, provided a basal age of 12,450 +/- 60 14C yr BP (Table 1; OS-7125; Thompson et al., 1999). This suggests that the nearby Bethlehem Moraine complex was deposited close to 12,000 14C yr BP (~ 14,000 cal yr). If the latter age is correct, the Littleton-Bethlehem Readvance occurred during the Older Dryas Chronozone as suggested by Thompson (1998) and Ridge et al. (1999). This cold interval began 12,200-12,000 BP and lasted only about 200 years (Donner, 1995; Wohlfarth, 1996).

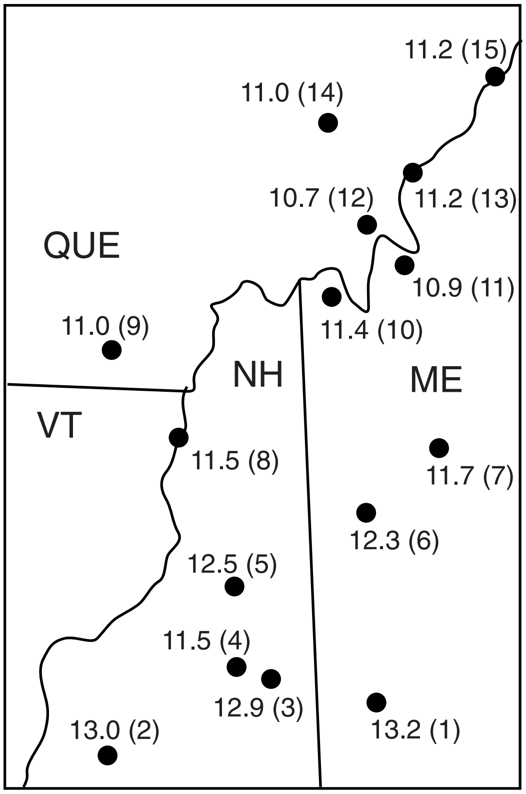

Figure 6: Map showing locations of ponds in the White Mountains and surrounding areas from which radiocarbon dates have been obtained from sediment cores. Only the oldest dates are shown, which are believed to approximate the time of deglaciation at each site. Numbers are ages in thousands of radiocarbon years BP, with site numbers (in parantheses) keyed to Table 1.

Radiocarbon ages of terrestrial organics from ponds in northern New Hampshire and neighboring Quebec indicate that the Boundary Mountains on the Quebec border were deglaciated as recently as 11,500 to 11,000 14C yr BP (Thompson et al., 1999). However, marine shell ages imply that the Champlain Sea transgression had already occurred by this time in the adjacent St. Lawrence Lowland to the north (Parent & Occhietti, 1988). If both the terrestrial and marine ages are correct, they require the development of an Appalachian ice mass that detached from the Laurentide Ice Sheet due to the marine incursion.

It is well known that a residual ice cap developed over northern Maine and adjacent Quebec, but this ice cap did not extend much farther southwest than Thetford Mines, Quebec (Parent & Occhietti, 1988). North of New Hampshire, moraines and other glacial deposits in the Sherbrooke area record the northwestward recession of the Laurentide ice margin (Gadd et al., 1972; Shilts, 1981; Dyke & Prest, 1987; Parent & Occhietti, 1988). Similarly, there is no indication that a late-glacial local ice mass formed over northern New Hampshire or adjacent Maine and Vermont. There are considerations that may help solve this problem. A correction for the marine “reservoir effect” or other factors could make the marine shell dates younger, and the terrestrial ages may be too young depending on how much they lag behind the time of deglaciation.

GLACIAL GEOLOGY OF THE PRESIDENTIAL RANGE

The name Goldthwait is synonymous with the glacial history of the Presidential Range. J.W. Goldthwait was the first to carry out an extensive study of glaciation in the Range (1913, 1916, 1938), where he reached three major conclusions: 1) the uplands above the cirques were eroded by both fluvial activity and continental glaciation, 2) the cirques were carved by alpine glaciers,

Table 1: Radiocarbon dates limiting time of deglaciation in the White Mountains.

Site No. and Name	Lab No.	Age, C-14 yrs BP	References
1 Cushman Pond	OS-7122	13,150+/- 50	Thompson et al. (1996)
2 Deer Lake Bog	QL-1133	13,000 +/-400	Davis et al. (1980)
3 Lost Pond	QL-985	12,870 +/-370	Davis et al. (1980)
4 Lakes of the Clouds	I-10684	11,530 +/- 420	Spear (1989)
5 Pond of Safety	OS-7125	12,450 +/- 60	Thompson et al. (1996)
6 Surplus Pond	OS-7119	12,250 +/- 55	Thompson et al. (1996)
7 Spencer Pond	AA-9506	11,665 +/- 85	C. Dorion (Unpublished)
8 Columbia Bridge	WIS-961	11,540 +/- 110	Miller & Thompson (1979)
9 Barnston Lake	GSC-420	11,020 +/- 330	McDonald (1968)
10 Lower Black Pond	OS-7123	11,500 +/- 50	Thompson et al. (1996)
11 Chain of Ponds	?	10,860 +/- 160	Borns & Calkin (1977)
12 Lac aux Araignees	GSC-1353	10,700 +/- 310	Shilts (1981)
13 Boundary Pond	GSC-1248	10,200 +/- 200	Shilts (1981)
14 Lac a la Truite	GSC-1289	11,000 +/- 240	Shilts (1981)
15 Lac Dufresne	GSC-1294	11,200 +/- 160	Shilts (1981)

as opposed to continental ice, stream erosion, or frost action, and 3) continental glaciation followed the last cirque glacier activity on the range (Davis, 1999). His evidence that cirque glaciers were not active following continental glaciation included: 1) the lack of looped end moraines on cirque floors, 2) till of a northern provenance on cirque floors, and 3) asymmetric cirque cross-valley profiles. Goldthwait (1913) did not support the concept that local glaciers extended far down valleys from an icecap centered on the Presidential Range, as proposed by Packard (1867a), Vose (1868), and Hitchcock (1878).

Over the next two decades, two workers disputed the conclusions of J.W. Goldthwait concerning the timing of continental and cirque glaciation in the Presidential Range. Johnson (1933) suggested the lack of end moraines in cirques is not sufficient evidence to conclude that continental ice post-dated cirque glacier activity in the Presidential Range, as he noted other alpine areas in the world that have never undergone continental glaciation but have cirques that lack moraines. Antevs (1932) sided with Johnson, concluding that Late Wisconsinan cirque glaciers existed in the Presidential Range; however, neither author provided a convincing explanation for the till of northern provenance on the cirque floors.

Richard P. Goldthwait (1939b, 1940, 1970) carried on his father's interest in the glacial history of the Presidential Range. In his 1939 and 1940 publications, he not only noted the observations of his father in support of cirque glacier activity predating the last overriding by continental ice, but he also observed roche moutonnees on cirque floors along with striae and grooves on cirque headwalls, which he believed could only have been formed by continental ice (Davis, 1999). Also, in his latter paper, he presented morphometric data on cirques and altitudinal estimates of firn lines for the former cirque glaciers in the Presidential Range (Fig. 7). From these data he calculated that, depending on the amount of winter precipitation, a 5 to 10°C mean summer temperature lowering would be necessary to support cirque glaciers in the Presidential Range today.

During the late 1950's, W.F. Thompson (1960, 1961) analyzed aerial photographs of the Presidential Range and refuted the Goldthwaits' view by arguing that the steep headwalls and sharp arêtes were indicative of active cirque glaciers following continental icesheet deglaciation. Although Thompson (1960, 1961) did not present field data to support his view, he believed that moraines of cirque glaciers had been obliterated by postglacial mass wasting processes (Davis, 1999). Work in Tuckerman Ravine during the late 1980's by D.J. Thompson (1999) suggests that a deposit consisting of large blocks believed to be a moraine by Antevs (1932) is a relict tongue-shaped rock glacier unrelated to cirque glacier activity.

Bradley (1981) challenged the Goldthwaits' view of the timing for cirque glaciation in the Presidential Range by noting that large boulders and diamicts at the mouths of north facing cirques were composed of lithologies derived from bedrock in the cirques. However, Gerath & Fowler (1982), Fowler (1984), Gerath et al. (1985), and Waitt & Davis (1988) examined the diamicts at the cirque mouths and concluded that the sediments are not till, but rather debris flow deposits.

The sides in the controversy concerning whether cirque glacier activity occurred pre- and/or post-continental glaciation can be summarized as follows. Proponents for post-icesheet local ice argue that the steep cirques and sharp arêtes are evidence for cirque glacier erosion following recession of continental ice. Opponents contend that this evidence is irrelevant because continental ice could have eroded the basins or because icesheet modification of older alpine forms was minor. Proponents interpret erratics drift derived from upslope sources in cirques as evidence for rebirth of local ice. Opponents counter that this evidence is weak because the till-like drift with the erratics was more probably deposited by postglacial mass wasting processes.

Opponents for post-icesheet local ice further support their views with the following evidence, which all suggest continental ice was last: 1) roche moutonnees indicating upvalley ice flow on cirque floors, 2) striae trending obliquely across cirque headwalls, 3) icesheet erratics in cirque-floor drift, and 4) apparent absence of moraines in and down valley from cirques. Proponents counter that cirque glaciers did not completely remove icesheet drift and that moraines in cirques are obscure because they are small, subdued by postglacial mass wasting, or thickly forested.

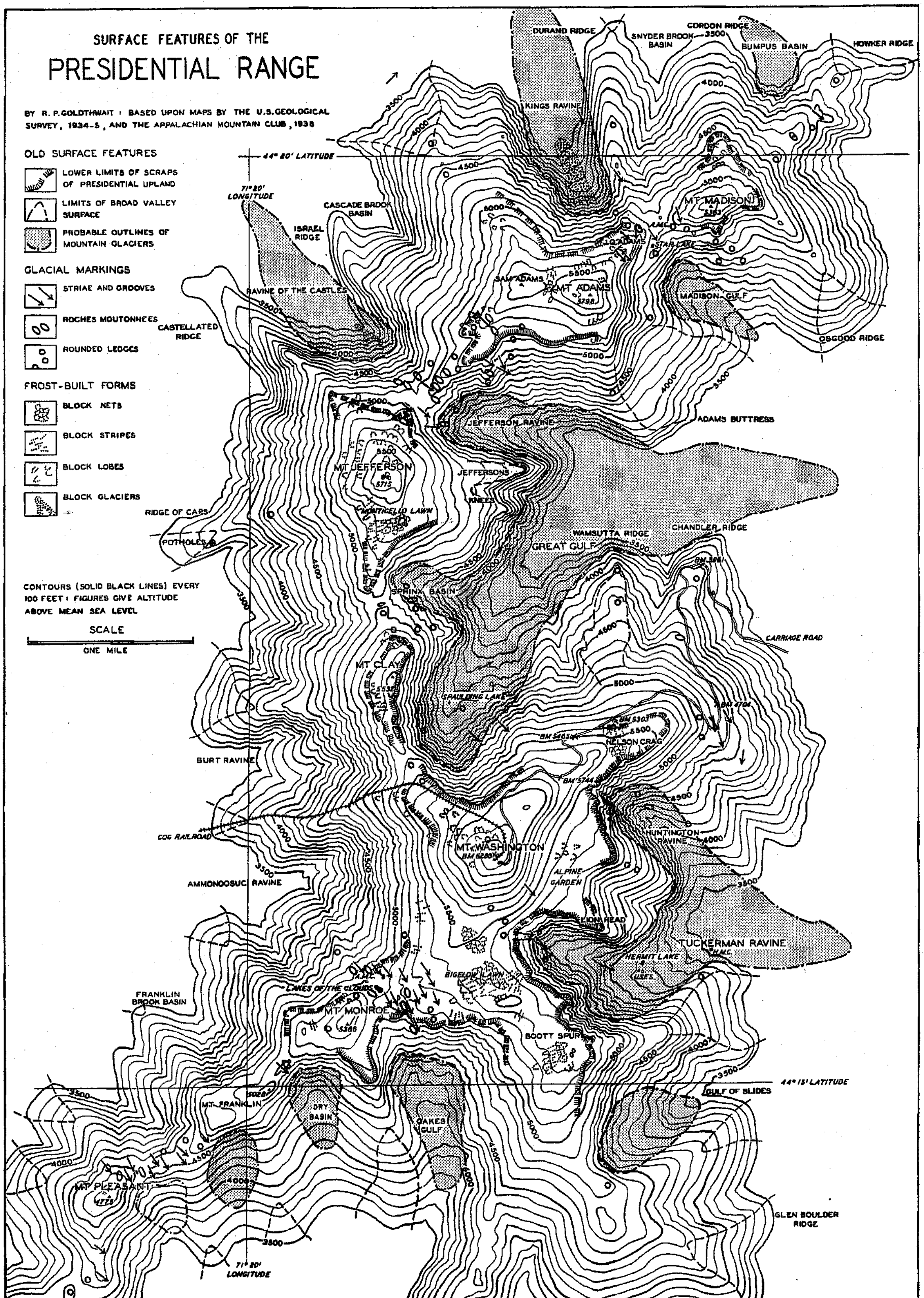

Figure 7: Surface features of the Presidential Range, including outlines of former cirque glaciers (from Goldthwait, 1970a).

An important theoretical argument against post-icesheet local ice in the Presidential Range is that, based on oxygen isotope and other proxies of late and immediate postglacial climate, such as pollen records, equilibrium-line altitudes rose rapidly to elevations well above cirque floors. Thus, cirque glaciers could not likely have been developed or sustained in the cirques following recession of continental ice.

Opportunities for developing a radiocarbon chronology for the deglaciation of cirques are limited because of the small number of tarns distributed across the Range. Spaulding Lake in the Great Gulf and Hermit Lake in Tuckerman Ravine, although shallow, may provide useful continuous postglacial records of sediment accumulation and should be cored.

Ponds near the cirques in the Presidential Range have provided minimum radiocarbon ages for continental ice retreat (Table 1; Davis et al., 1980; Spear, 1989; Spear et al., 1994). Organic material from sediment at the base of a core retrieved from Lost Pond at an elevation 650 m in Pinkham Notch on the east side of the Presidential Range provided a radiocarbon age of 12,870+/- 370 yrs BP (QL-985; Spear et al., 1994). Organic material from sediments near the base of cores taken from the lower of the two Lakes of the Clouds at an elevation of 1542 m in the alpine zone between Mounts Monroe and Washington yielded a radiocarbon age of 11,530+/-420 yrs BP (I-10684; Spear, 1989). Pollen data from sediments below the radiocarbon-dated level in the lower Lakes of the Clouds site correlate with the tundra pollen zone from Deer Lake Bog at an elevation of 1300 m on Mount Moosilauke, which provides a radiocarbon age of 13,000+/-400 yrs BP (QL-1133; Davis et al., 1980).

Given the model that continental ice thinned, separated, and retreated northward from the mountains of northern New England during Late Wisconsinan deglaciation (Goldthwait & Mickelson, 1982; Hughes et al., 1985; Stone & Borns, 1986; Borns, 1987; Davis & Jacobson, 1985; Thompson & Fowler, 1989), this entire process appears to have been very rapid. If these radiocarbon ages are taken at face value, they require almost 900 m of continental ice thinning in less than a few hundred years, a circumstance requiring very rapid warming of the local climate, warming too rapid to support local cirque glaciers. These ideas on the surficial geology of the White Mountains have been discussed on previous society field trips in the area (Davis et al., 1988; Davis et al., 1993; Davis et al., 1996).

Current work designed to refine the deglaciation chronology for the Presidential Range uses cosmogenic radionuclides 10 Be and 26 Al produced in quartz from boulders and bedrock. These exposure dating techniques (Bierman, 1994) may not provide the temporal resolution of AMS radiocarbon dating, but the method does allow samples to be collected from sites where radiocarbon-datable materials are not available. As a test of the thinning continental ice model for deglaciation of the Presidential Range, a suite of bedrock and boulder samples with quartz veins were collected on an altitudinal transect from the summit of Mount Washington to the floor of Pinkham Notch near Lost Pond for cosmogenic nuclide dating. The abundance of 10Be and 26Al in frost-riven bedrock samples from near the summit area of Mt. Washington is much higher (1.5 to 8 times) than expected had the peaks been covered by active, erosive ice during the Late Wisconsinan maximum (Bierman & Davis, 2000). Two samples provide 10Be model ages of 124,000 and 22,000 years BP. However, a sample from one of the large boulders on the tongue-shaped rock glacier on the floor of Tuckerman Ravine (D.J. Thompson, 1999) appears to be late-glacial in age (Bierman & Davis, 2000). The mountain-top samples are consistent with two different scenarios, both of which have significant implications for understanding the spatial and temporal patterns of glaciation and glacial erosion in northern New England: 1) late Wisconsinan continental ice was thinner than previously supposed, leaving Mt. Washington's summit exposed since marine isotope stage 6, or 2) the summit was covered by glacial ice during the late Wisconsinan, but the ice was thin enough to be frozen to its bed (< 100 meters thick). Thus, the cold-based ice was unable to erode much rock, allowing cosmogenic nuclides to be inherited from prior periods of exposure. In either case, continental ice in New England during the late Wisconsinan was thinner than previously believed, consistent with low basal shear stresses and/or the presence of active ice streams in the notches.

FIELD STOPS

An excellent overview of the White Mountains is provided by the Mount Washington 1:100,000-scale topographic map. An excellent large-scale map for stops in Pinkham Notch is the 1:20,000-scale map of *Mount Washington and the Heart of the Presidential Range, New Hampshire* (Washburn, 1989); the large-scale quadrangles covering other field trip stops are listed under the stop descriptions. You may also wish to refer to the *New Hampshire Atlas,* published by the DeLorme Mapping Co. Your driving distances may vary slightly from the mileage given here, depending on the route taken in some of the turnarounds, larger gravel pits, etc. Note: Some of the stops are located on private property. Permission must be obtained from the owners for any future visits!

Saturday, November 3

0.0 Mileage for the road log through Pinkham Notch begins at Stop 1, the Great Gulf Wilderness Parking Area on the west side of NH 16, about 6 miles south of its intersection with US 2 in Gorham, or about 3 miles north of Wildcat Ski Area. Walk down the trail and across the footbridge over the Peabody River; we will return to the outcrops below the bridge later (Stop 1B). About 150 feet beyond the footbridge, the old Great Gulf Link Trail enters sharply from the left; follow this abandoned trail approximately 500 feet north (downstream) to some rock ledges on the west bank of the Peabody River (Stop 1A).

STOP 1A. Schists of the Rangeley Formation, Peabody River, consist of coarse-grained schists with some interbedded quartzites and granofels. Calc-silicate pods occur in groups along bedding planes. Sillimanite nodules (pseudomorphs after andalusite) and un-oriented spangles of retrograde muscovite give the bedding/foliation surfaces a knobby texture. Whereas there are some bed-to-bed variations, the major element, trace element, and stable isotope compositions are all consistent with a metamorphosed shale. There is no evidence that these rocks reached the sillimanite-potassium feldspar (muscovite absent) grade. Return to the outcrops under the footbridge.

STOP 1B. Migmatites of the Rangeley Formation, Peabody River. The migmatite leucosomes are usually elongate blebs or stringers, rather than continuous layers, and the intensity of migmatization is highly variable. The age of this migmatization is circa 404 Ma. Calc-silicate pods and blocks with folded layering are abundant, and usually lie parallel to the migmatite layering, although sometimes they are at an angle to it. These rocks are obviously quite heterogeneous, but in bulk they are geochemically similar to the protolith schists at Stop 1A (Allen, 1992, 1996b). This suggests that the migmatization of these rocks is the result of metamorphism in an iso-chemical or closed system. The migmatites have a lower oxygen isotope value than the schists and exhibit open-system isotopic behavior (Allen, 1992, 1996b), which suggests infiltrating fluids may have been important in driving the migmatization process. Isotopic fractionation between the melanosome and leucosome components of the migmatites may be the result of partial melting. These migmatites have an identical metamorphic mineral assemblage to that at Stop 1A (sillimanite-muscovite), but garnet-biotite geothermometry indicates temperatures were higher (Allen, 1992, 1996b). Two generations of pegmatite cut across the outcrop.

STOP 1C. Early and late fold structures within the unmigmatized Rangeley Formation, Peabody River. Proceed south to the junction with the Great Gulf Link Trail. Continue 1,700 ft (550 m) south along the Great Gulf Trail following the West Branch of the Peabody River upstream to where the marked ski trail diverges off to the right (west) at about elevation 1,450 feet. Scramble down to the river outcrops where well bedded schists and quartzites of the Rangeley Formation show bedding, S0, and S1 schistosity nearly perpendicular to it, suggesting minor F1 folds are present. Exposed in the stream is a partial F1 syncline showing S0, defined by schist, quartzite, and garnet coticule layers, folding around. Well developed S1 axial planar schistosity is present. The coticule layers show excellent minor fold asymmetry, consistent with the larger fold structure. F4 crenulations with axial surfaces striking northerly and dipping east are overprinted by F5 crenulations with axial surfaces striking easterly and dipping north. The F4/F5 intersection pattern is best seen on the schist bedding plane and S1 foliation surfaces and most often in loose blocks in the stream bed

If we have time, continue upstream another 700 feet (220 m) to elevation 1475 feet. Along the way observe in the schist layers the aligned pseudomorphs of andalusite, L1. The lineation is parallel to the F1 hinge lines throughout the Range. Sometimes the lineations show relict chiastolite crosses and always it is easy to see the replacement by sillimanite and muscovite. At the end of the traverse there is a well developed F1 minor fold showing east-northeast facing direction, consistent with facing directions throughout the Presidential Range that are easterly.

Bushwack back down the river, return to the Great Gulf Trail and the parking lot. Return to vehicles and drive south on NH 16, towards Wildcat Ski Area and Jackson, NH.

0.3 Turn left into 19-Mile Brook trailhead parking area. Walk across the highway bridge over 19-Mile Brook, cross NH 16 (watch for traffic!), and head down a gated access trail along the brook.

STOP 2. Rangeley migmatites with calc-silicate pods, 19-Mile Brook and Peabody River. Exposed here are extensive washed and pot-holed outcrops of gray to rusty orange migmatitic gneiss. The rocks are generally gray adjacent to granite/pegmatite intrusions, rusty elsewhere. Locally the gneiss appears to grade into granite. There are diffuse pegmatite bodies, as well as pegmatites with distinct sharp contacts. Pods with moats appear to occur in horizons, generally parallel to the layering. A nice polished surface on the floor of a pool in 19-Mile Brook, approximately 100 feet upstream from its confluence with the Peabody River, shows leucosome/melanosome layering in the migmatite gneiss and the internal structure of pods and their relationship to one another (a family of pods within a larger pod!). Return to vehicles and continue south on NH 16.

3.3 Park in the upper parking lot on the left, opposite the entrance to Mount Washington Auto.

STOP 3. Cirques of the northern Presidential Range, Glen House Site. Mount Washington (elev. 1950 meters a.s.l.), is the highest peak in the Presidential Range and the highest peak east of the Mississippi River and north of the Carolinas. Along with its superior elevation, Mt. Washington also boasts the world's worst weather, its famous Observatory having recorded, along with other extremes of cloudiness, temperature, and precipitation, the world's highest wind gust of 231 mph in 1934. Time constraints and early winter on the mountain do not permit us to visit the summit, but if the weather permits, fine panoramic views of the Presidential Range and the great glacial cirques cut into its eastern slopes are visible from this location.

We know that all of the peaks here were overridden at least once by continental ice because lodgement till has been found near the summit of Mt. Washington at an elevation of 1,900 meters. Most workers in this area agree that the last continental icesheet (the Late Wisconsinan Laurentide Icesheet) arrived here about 25,000 years ago, reached its peak about 18,000 years ago, and had fully retreated from lowlands in the area by about 12,000 years ago. All ages provided in the surficial geological stops of this field trip will be given in radiocarbon years Thompson et al. (1999) have provided a detailed review of radiocarbon carbon ages for the region north and northwest of the notches as summarized in the Table 1. However, Ridge et al. (1999) question whether many of these radiocarbon ages are accurate, and propose that deglaciation in the White Mountains was up to hundreds of years more recent than the oldest ages provided in Table 1.

Although on face value, the radiocarbon ages in Table 1 might suggest that deglaciation and organic sediment accumulation occurred earlier in bogs and ponds at lower elevations (ex. Lost Pond) than at higher elevations (ex. Lakes of the Clouds). However, palynological data (Davis et al., 1980; Spear, 1989; Spear et al., 1994; Cwynar & Spear, 2001) suggest that postglacial revegetation occurred at nearly all elevations at roughly the same time. Hence, the pollen data suggest that regional climate warmed rapidly during the early postglacial, which supports the argument presented earlier that glacial equilibrium-line altitudes rose too quickly to allow cirque glaciers to exist during the postglacial. Return to vehicles and continue south on NH 16.

1.6 Paved parking area on the right. The road cuts opposite are the destination of a traverse made in Stop 4B.
2.0 Pull off into a paved parking area on the right hand side of the road. Park near the upper end. Make your way down to Emerald Pool and then right over ledges and through woods to a small beach just upstream of the pool.

STOP 4A. Emerald Pool, Peabody River. Billings (1941; Billings & Fowler-Billings, 1975) and later Hatch (Hatch & Moench, 1984; Hatch & Wall, 1986) mapped these rocks as part of a belt of what is now called the Madrid, Smalls Falls, and Littleton, exposed in the West Branch of the Peabody River. Allen (1992, 1996a,b) concurred that the entire Siluro-Devonian stratigraphic sequence was exposed here, but interpreted them to be in-folded by a north-northeast trending doubly-plunging syncline within the Rangeley migmatites, correlated with similar occurrences on Mount Moriah to the northeast (Fig. 2; Allen, 1996a). Eusden suggests that the gneissic rocks mapped at Emerald Pool are all migmatized Rangeley Formation. In this interpretation, the rusty schist and calc-silicate horizons at Emerald Pool are members of the Rangeley and not the Madrid and Smalls Falls Formations, and no Littleton is present. This is based on observations of similar Rangeley members elsewhere in the Presidential Range and, as in Allen's interpretation, detailed mapping surrounding Emerald Pool.

Head west into the woods to a small stream and follow that up to elevation 1675 feet where a 3 m by 10 m patch calc-silicate granofels is completely surrounded by migmatite. Is this another block of calc-silicate within the Rangeley or is it the Madrid Formation? Return to the parking area and carefully cross the highway to roadcuts on the east side.

STOP 4B. Littleton to Rangeley section, or Rangeley section only? NH 16. Walk north along this half-mile long series of outcrops on the east side of NH 16. Allen (1992, 1996a,b) proposed that these outcrops represent a succession from gray, "sinewy" migmatites, lacking calc-silicate pods, of the lower Littleton rocks back to the rusty of the Rangeley Formation, with distinctive pods and exotic cobbles. An outcrop of clean quartzite and schist just off the road in the woods between these roadcuts may be Perry Mountain. At the south end of the second large highway outcrop, north of the Emerald pool parking lot, there is more migmatite, now with calc-silicate pods and a rare Mesozoic (?) basalt dike cutting through. Allen (1992, 1996a,b) interpreted these rocks, together with those above Emerald Pool, as an in-folded septum of Smalls Falls, Madrid and Littleton rocks within the migmatites of the Rangeley Formation. Eusden would interpret all of this section as the Rangeley Formation that has several different lithologic members. Return to vehicles and continue south on NH 16.

2.3 Pull off into a paved parking area on the right hand side of the road. Park near the lower (northern) end, where access to pavement outcrops in the Peabody River is obvious.

STOP 5. Wildcat Granite, Peabody River. For those familiar with the controversy, granitization rears its ugly head! This is the type locality for the rock Allen (1992, 1996a,b) called Wildcat Granite. There are clearly two different phases: "G" consists of medium grained, whitish-weathering two-mica granite; "R" is much coarser grained, orangish-weathering granofels, also bearing both muscovite and biotite, but with much more abundant biotite. R contains abundant calc-silicate pods, identical to those found in the metasediments, rimmed by strong reaction zones, and its textures and mineralogy suggest that it may be formed from completely melted and recrystallized Rangeley schists. Eusden would retain the name Rangeley Formation migmatite for Allen's R Wildcat Granite. This reflects a minor problem of semantics in determining, while in the field, when a migmatite, with continued melting, becomes a granite. Both the G and R phases are extensively intermingled in a complex fashion, with wispy biotite-rich schlieren observed throughout the outcrop. Geochemically, the G phase is characteristic of an S-type, while the R phase has a composition intermediate between that of the G phase and those of the schists and migmatites (Allen, 1992, 1996b). Second-generation, white, tourmaline-bearing pegmatites crosscut both phases.

3.1 Turn left into Wildcat Ski Area.

STOP 6. Pinkham Notch and the cirques on Mount Washington. We are at approximately the height of land in Pinkham Notch, at an elevation of about 650 meters a.s.l., at the eastern base of Mount Washington (1950 m a.s.l.). From here, we have a good view of Tuckerman Ravine, to the left, and Huntington Ravine, to the right.

An interesting side trip from this stop is the hike on the Thompson Falls Trail to its end (0.8 miles) at ledges of Wildcat Granite grading into Rangeley migmatites. The contact between the Wildcat Granite and the surrounding migmatitic metasedimentary rocks is gradational—not a sharp intrusive contact. On a good day, excellent views of Mt. Washington and its ravines can be had from these falls.

4.0 Turn right into the Appalachian Mountain Club's (AMC) Pinkham Notch Base Camp. Parking may be difficult to find, in which case continue to the auxiliary parking lot on the south side of the Cutler River.

STOP 7. Crystal Cascade, Cutler River. Follow the Tuckerman Ravine Trail 0.3 miles up the Cutler River. On the way up, just before the first bridge over the river, the trail crosses a dry stream bed that used to be the course of the Cutler River. Interestingly, this course would take the Cutler north to the Androscoggin River drainage, whereas the stream today flows south to the Saco River drainage. A subtle shift in fairly recent times (Holocene?) changed the river's course dramatically. The rocks at the Cascade are Mesozoic volcanic vent agglomerates (Billings, 1979), exposed in the wall at the viewing area. A complete upright section from Littleton (upstream of the falls) to Madrid (at the falls), Smalls Falls, and unmigmatized Rangeley (downstream of the falls) as well as exposures of granite and pegmatite can be seen in the Cutler River. Viewing the rocks at and above the Cascade is difficult due to the steep terrain, however, a traverse downstream of the first bridge through the Smalls Falls-Rangeley contact is quite possible.

Several other interesting short hikes to sites of geological interest may be made from the AMC's Pinkham Notch Base Camp, including Square Ledge (0.5 miles) and Lowe's Bald Spot (1.8 miles). To get to Square Ledge, cross to the east side of NH 16 carefully. Follow the Square Ledge Trail 0.5 miles to the top of Square Ledge, an outcropping of non-descript gray Rangeley migmatites with an excellent overlook of Pinkham Notch and view of Mt. Washington and its ravines.

To get to Lowe's Bald Spot, follow the Old Jackson Road (trail) 1.8 miles to the Mount Washington Auto Road, roughly following the contact between the Smalls Falls and Madrid or Littleton Formations. At the north end of the trail before the junction with the Auto Road, the Madrid and Smalls Falls Formations are missing from the stratigraphy, we think cut out by a fault. In this region the Littleton is juxtaposed against the Rangeley. We have called this discontinuity the Pinkham Notch Normal Fault and it extends for about 2 km to the northeast crossing the Auto Road twice. We have also recognized an identical discontinuity, the Graham Trail Normal Fault, about 2.5 miles to the south along strike in the New River drainage basin. Both of these faults are interpreted to be pre-metamorphic structures occurring during deposition in a slope-rise environment, however, lack of kinematic indicators make other interpretations viable; eg. thrust or strike-slip.

30 m uphill from the 1-mile marker on the Auto Road at an elevation of 2050 ft there is a small brook. Follow this upstream about 100 m to the discontinuity between the Silurian Rangeley Formation and Devonian Littleton Formation. As is typical in the Central Maine Terrane, due to available exposure you can not actually put your hand on the Pinkham Notch Normal Fault. However, you can get to within about 10 meters between outcrops of Littleton and Rangeley. The outcrop pattern of the fault is complicated by F1 folding. The interpreted footwall rocks of the Rangeley Formation are folded by an F1 syncline forming a small hill (elev. 2,250 ft.) and the Littleton in the hangingwall is folded by a F1 anticline into a "window".

At Lowe's Bald Spot, schist and quartzite of the Littleton can be observed with tops up. The dip of bedding, S0, is northerly, near the hinge of a macroscale F4 fold, the Lowe's Bald Spot Syncline. Thin layers of discontinuous garnet plus quartz or coticule are seen. F4 and F5 crenulations are seen intersecting each other. These outcrops are at the western limit of F5 deformation. Just west of Lowe's Bald Spot only F4 deformation is seen. This pattern of spatially restricted phases of deformation is seen throughout the Presidential Range.

4.5 Pull off on right shoulder of NH 16, at a small waterfall opposite a turnout on the northbound side and just before a turnout on the southbound side

STOP 8. Smalls Falls, Madrid and Littleton Formations, Pinkham Notch. The rocks at the base of the falls are laminated rusty schists of the Smalls Falls Formation. These are overlain by layered green calc-silicate granofels of the Madrid Formation near the top of the falls, and above that by well bedded aluminous schists of the Littleton Formation. Across the road along the Glen Ellis River are stream outcrops of Rangeley Formation migmatite.

Another interesting side trip is the Glen Ellis Falls Scenic Area, just ahead. Follow the path under NH 16 and down to falls. There's a plaque above the Falls titled "Geology of Glen Ellis Falls" and it says on it that "The Ellis River which had flowed uninterrupted during pre glacial times was forced by violence and the struggle of the landmasses to plunge over the headwall of a glacial cirque." Phooey! There's no cirque here and if you remember STOP 7 below Crystal Cascade, the Cutler River only recently adopted its present course. Glen Ellis Falls is here because a major silicified zone, likely a Mesozoic fault with only a few meters of offset, juxtaposes the calc-silicate pod-bearing Wildcat Granite or Rangeley Formation migmatite against the Smalls Falls Formation. The 3 meter wide zone of massive quartz with many cross cutting quartz veins is quite resistant and forms the escarpment over which the Falls plunge. The silicified zone can be followed for about 1 mile in total, up and downstream of the Falls, and in a few places you can find tiny vugs containing 0.5 cm long quartz crystals.

5.2 The large road cut on the west side of the NH 16, opposite a turnout on the east side overlooking the Ellis River valley and the town of Jackson, is assigned to the Wildcat Granite.
13.2 Turn left on NH 16A into the village of Jackson.
13.6 Turn left on NH 16B at The Wentworth.
13.9 Park on wide right (east) shoulder at Jackson Falls picnic area. Walk to broad exposures in Wildcat Brook.

STOP 9. Albany Porphyritic Quartz Syenite of Ring Dike E2, Jackson. (Jackson quadrangle) NO HAMMERS PLEASE! The Albany Porphyritic Quartz Syenite in composite ring dike E3. This is one of the best (and most scenic) localities to examine the Albany—about 2000 ft of continuous exposure is present between here and Jackson along Wildcat Brook. Just downstream of the iron bridge, a screen of Siluro-Devonian gneisses and granite 110 ft (33 m) wide is intruded by the Albany. Downstream from this screen, the Albany is relatively uniform in mineralogy and texture and contains a few small (2-5 cm) inclusions. Upstream, large (up to 1 m) inclusions of feldspar-poorer porphyritic syenite are enclosed by the Albany. Turn vehicles around, and return to NH 16.

Sunday, November 4

0.00 The Mileage for the road log through Crawford and Franconia Notches begins at the traffic light on north Main Street in Gorham, at its intersection with US 2 West. Go west on US 2.
9.65 Turn left onto Valley Rd.
10.05 Turn left and drive into the Corrigan Pit.
10.15 Park in pit area (circle around if possible and keep cars in line for easy exit).

STOP 10. Corrigan Pit, Randolph. (Mt. Washington 7.5 x 15-minute quadrangle) The Corrigan Pit is located next to the Israel River, which flows northwest and joins the Connecticut River at Lancaster. A short distance east of here, at Bowman, US 2 crosses the 457-m divide into the Androscoggin River basin. As the late Wisconsinan glacier margin receded down the Israel Valley, the Bowman Stage of glacial Lake Israel was impounded in this area and spilled eastward across the divide. Glaciolacustrine fans and deltas deposited in this lake have been exposed in several pits in Randolph and Jefferson. An ice-contact delta graded to the Bowman Stage forms a narrow terrace along US 2 just north of the Corrigan Pit.

Northwest of here, meltwater channels were cut into the west side of Boy Mountain along the margin of the retreating Israel Valley ice tongue (Lougee, n.d., 1939; Goldthwait & Mickelson, 1982). Lateral meltwater channels and ice-contact deposits also exist on the south side of the valley (Crosby, 1934; Goldthwait & Mickelson, 1982; Thompson et al., 1999). Lougee (n.d.) noted esker and lake-bottom sediments downvalley from here.

As we consider the glacial features in the Corrigan Pit and surrounding area, the principal question concerns the mode of deglaciation. By analogy with Alaskan glaciers, Goldthwait & Mickelson (1982) interpreted this area as demonstrating the early stagnation of the late Wisconsinan ice sheet, with development of a ragged ice margin as the glacier thinned over high-relief terrain. In the present authors' opinion, it is more likely that the receding ice remained active. It built several prominent end moraines in the upper portion of the Israel valley and numerous small moraine ridges near Jefferson village.

The Corrigan Pit is located in a low moraine ridge with a surface elevation of up to 439 m. The sections in the southern part of the pit, close to the river, have exposed a sequence consisting of glaciolacustrine sand and gravel which coarsens upward and is overlain by stony glacial diamict. Much deformation is evident throughout the section. The three principal stratigraphic units are as follows: (1) The lowest unit consists of glaciolacustrine sand (base not exposed) with planar foreset beds dipping NE to E. In places there is much faulting in the central to upper parts of this unit, which probably resulted from ice shove. (2) The lacustrine sand is overlain by a variable thickness (approximately 2-3 m) of ice-proximal gravel and sand. The gravel fraction is poorly sorted and angular. This unit contains scattered outsize boulders and diamict lenses (flowtills). Locally abundant shear structures indicate ice shove to the northeast. (3) The uppermost unit consists of up to 6 m of bouldery, silty-sandy glacial diamict containing abundant lenses of washed sediment. This unit appears identical with the regional late Wisconsinan surface till. Some of the sand lenses in the diamict are highly deformed and suggestive of ice shove.

The deposits in the Corrigan Pit are believed to record the eastward advance of a glacier margin into a lake ponded in the Israel River valley. The constructional morainic topography suggests that these deposits resulted from a readvance into the Bowman stage of glacial Lake Israel during the overall recession of the ice sheet.

Turn right out of pit and return to US 2.
10.60 Turn left onto US 2 (west) toward Jefferson.
13.35 Just before large yellow house on right, note meltwater channel crossing US 2. This is one of a series of channels on the side of Boy Mountain noted by Lougee (n.d.; 1939) and Goldthwait & Mickelson (1982).
13.85 Turn left (south) onto NH 115.
17.95 Note sign on left describing the famous Cherry Mountain landslide of 1885 (see Gosselin, 1985). The slide path crosses the road just downhill from here (back toward Jefferson), but is hardly discernible anymore.
23.85 Turn right onto US 3 (north).
23.95 Turn left into the Twin Mountain Sand & Gravel Pit.

STOP 11. Twin Mountain Sand & Gravel Pit, Carroll. (Bethlehem 7.5 x 15-minute quadrangle) The large pit at this stop is located in an ice-contact delta, known as “the Carroll delta” (Lougee, 1940), which built southward into glacial Lake Ammonoosuc. This deposit formed when the receding late Wisconsinan ice margin was pinned against Beech Hill to the west and Cherry Mountain to the east. Figs. 4, 5 and 8 show the relationship of the Carroll delta to meltwater channels, inferred ice-margin positions, and other features associated with the glacial lake.

Lougee (1940) described this area in detail. He noted the fluvial incision of the delta resulting from a drop in lake level when recession of an ice tongue farther down the Ammonoosuc Valley opened up lower spillways for Lake Ammonoosuc. The Alder Brook channel that was cut into the west side of the delta as the lake fell is clearly seen along the railroad track. The water that discharged through this channel came from the ice-margin positions shown in Fig. 8, and perhaps later from glacial Lake Carroll just north of the delta.

Aerial photographs from 1955 (prior to the pit operation) show meltwater channels trending north to south across the delta top. These photos also show a steep ice-contact slope on the north edge of the delta, where the processing plant is now located. Sediment was supplied to the delta both from meltwater channels on the hillside to the northeast, and directly from the adjacent ice margin via subglacial drainage (marked by an esker along the railroad track north of the pit). The delta top has an elevation of approximately 450 m (1475 ft). It was graded to the Gale River 2 stage of glacial Lake Ammonoosuc, with a spillway at about 445 m (1460 ft). This outlet channel can be seen along US 3 southwest of Twin Mountain village (Fig. 4). Lower surfaces resulting from downcutting of the Carroll delta will be examined at Stop 12.

Remnants of coarse gravel forming the delta topset beds can be seen along the upper east wall of the pit. In 2000 the sandy foreset beds were best exposed in a newly deepened area east of the processing plant. This opening showed many thrust faults in the delta foresets, resulting from ice shove against the back side of the delta, and providing further evidence of late-glacial ice activity in the region. The Beech Hill moraines just northwest of here (“B” on Fig. 4) are contemporaneous with the Bethlehem Moraine complex to the west and moraines in the upper Israel River valley to the east (Thompson et al., 1999).

24.00 Leave pit and turn right (south) on US 3.
26.10 Turn right (west) onto US 302 at Twin Mountain village.
27.10 Turn right and follow road into N.H. Dept. of Transportation pit complex.
27.85 Park in the northeast end of pit area.

STOP 12. D.O.T. pit, Carroll. (Bethlehem 7.5 x 15-minute quadrangle) This pit is cut into sand and gravel deposits with graded upper surfaces at about 435 m and 423 m (Fig. 8). The best recent exposures have been in the northern part of the pit complex, which is mostly in the 435-m terrace. Boulders on the pit floor, together with the shallow depth of the workings, suggest that till may occur just beneath the floor. Small outcrops of granite gneiss occur elsewhere in the floor of the pit complex.

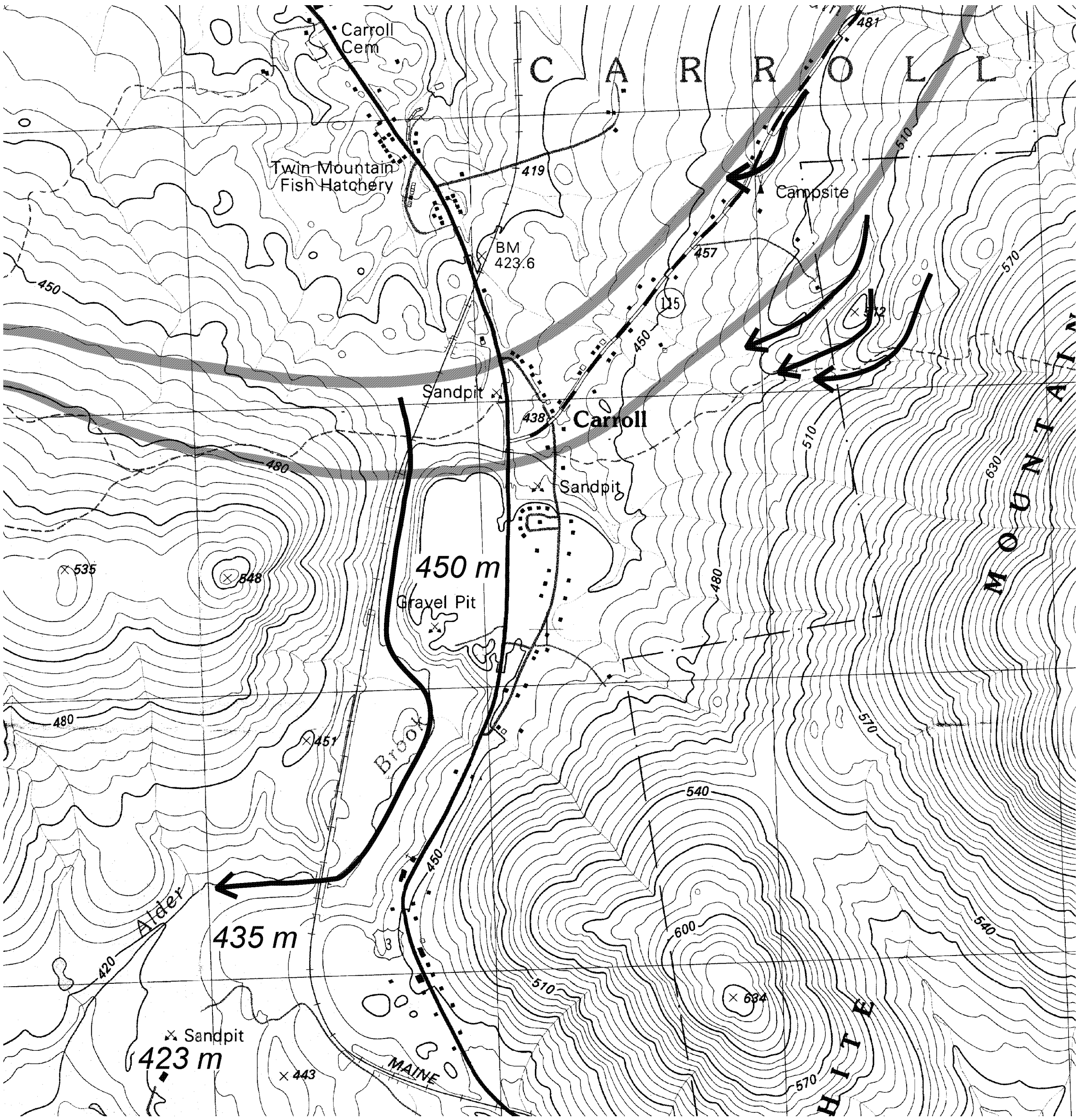

Figure 8: Part of the Bethlehem 1:25,000 metric quadrangle, showing the location of Stops 11 and 12; the ice-contact Carroll delta at elevation of 450 m; the 435 and 423-m surfaces graded to lower stages of glacial Lake Ammonoosuc; inferred ice-margin positions (gray lines); and meltwater channels (arrows).

The fluvial sediments seen here were deposited along the southern part of the Alder Brook channel. They probably were derived at least partly from downcutting of the Carroll delta to the north, as described under Stop 10. The abundant gravel suggests that meltwater may have continued to issue from the northern ice margin position shown in Fig. 8 when the 435-m surface was formed. This terrace was graded to the short-lived Gale River 3 Stage of glacial Lake Ammonoosuc, which drained through the G3 spillway channel located 5 km west-southwest of here (Fig. 4).

The lower (423 m) terrace is exposed near the garage at the entrance to the pit area, where a section shows several meters of sandy delta foreset beds dipping SSW. No topset beds are present here. These deltaic sediments may be the remnant of a higher and earlier delta that has been trimmed, or they may have been deposited as part of a new delta when drainage from the Alder Brook channel entered the Gale River 4 Stage of Lake Ammonoosuc. The 423-m surface correlates with the G4 spillway, which is a prominent channel just west of the G3 outlet (Fig. 4).

Return to US 302 and head east back towards Twin Mountain.
29.7 Go straight through on US 302 east at the intersection with US 3.
34.9 The Mount Washington Hotel at Bretton Woods, on the left, with a nice view of Mount Washington.
38.3 The AMC's Crawford Notch Hostel. An interesting side trip is the hike to the summit of Mount Willard (1.4 miles), where Conway Granite is exposed, and which provides a magnificent vista of U-shaped Crawford Notch to the south.
38.6 Park on southwest side of US 302, near the head of Crawford Notch ("the gate of the notch").

STOP 13. "The Gateway" at the head of Crawford Notch. (Crawford Notch quadrangle) We will walk east along the railroad tracks to a spectacular view down Crawford Notch. The steep slope at the head of the notch is not a cirque headwall at this low altitude, just as the precipice at Glen Ellis Falls in Pinkham Notch is not a cirque headwall. This valley knickpoint probably predates Quaternary glaciation, and its presence here may result from harder bedrock types, such as the hornfels of Silurian Rangeley gneiss in the contact aureole of an intrusion of Jurassic Conway Granite, at this location. . The darker crest of the ridge on Mount Webster, on the left of east side, is also Silurian paragneiss and marks the roof of the Conway Granite intrusion. This contact is exposed at road level on US 302 approximately 0.9 miles east from the head of the notch.

Both walls of Crawford Notch (Mt. Webster, and the Willey Range on the right, or west side) are heavily scarred by landslides. The largest landslide scar on the west side of the notch occurs on Mt. Willey, the site of the infamous Willey slide that occurred at 3 a.m. on Monday, August 28, 1826. As the story goes, upon hearing the roar of the slide, all family members ran out of their house, only to be obliterated while the house survived. Apparently a large boulder sitting directly upslope split the slide into two paths around the house. Following a hot and dry summer, landslides occurred all over the White Mountains during this two-day storm in 1826 (Silliman et al, 1829; Ramsey, 1988).

The head of Crawford Notch also was a spillway for glacial Lake Crawford to the west, which was contemporaneous with glacial Lake Pigwacket in the Conway area south of Crawford Notch (Thompson, 1997). The voluminous glacial meltwater that flowed down Crawford Notch during late-glacial time was orders of magnitude greater than postglacial runoff, and was responsible for the smaller V-shaped incision superposed on the classic U-shaped trough.

Turn vehicles around and return west on US 302.
47.5 Turn left onto US 3 at Twin Mountain village, south towards Franconia Notch
58.1 US 3 joins Interstate 93, which soon becomes the Franconia Notch Parkway.
60.1 Take Parkway Exit 2 for "Old Man Viewing". Turn right at the bottom of the ramp, then left into the Old Man parking lot and park. Walk about 1/3 of a mile to the Old Man Viewpoint on the north shore of Profile Lake.

STOP 14A. Old Man Of The Mountain, Franconia Notch. (Franconia quadrangle) The Old Man of the Mountain was made famous by Nathaniel Hawthorne's classic work, "The Great Stone Face." Most of the material below is summarized from Fowler (1982), which details the structural geology and rock mechanics that account for the delicate stability of this remarkable rock formation.

The "Old Man of the Mountain," New Hampshire's most famous landmark and the emblem on its State Seal, license plates, and recently minted millennial quarter, consists of seven large blocks of Conway granite bedrock juxtaposed in such a way as to provide the profile seen from Profile Lake (Fig. 9).

The Profile is made up of six fortuitously-broken, plate-shaped blocks of Conway Granite positioned above each other in such a way as to create the distinctly "human profile" best seen from the shore of Profile Lake, 550 meters below. The Profile is approximately 21 meters high and is estimated to weigh between 10,000 and 13,000 tons. The six blocks comprising the Profile represent, from top to bottom: 1. forehead, 2. eyebrow, 3. nose, 4. base of nose and upper lip, 5. lower lip and chin, portion separated from the cliff, and 6, lower lip and chin, portion behind No. 5 and not separated.

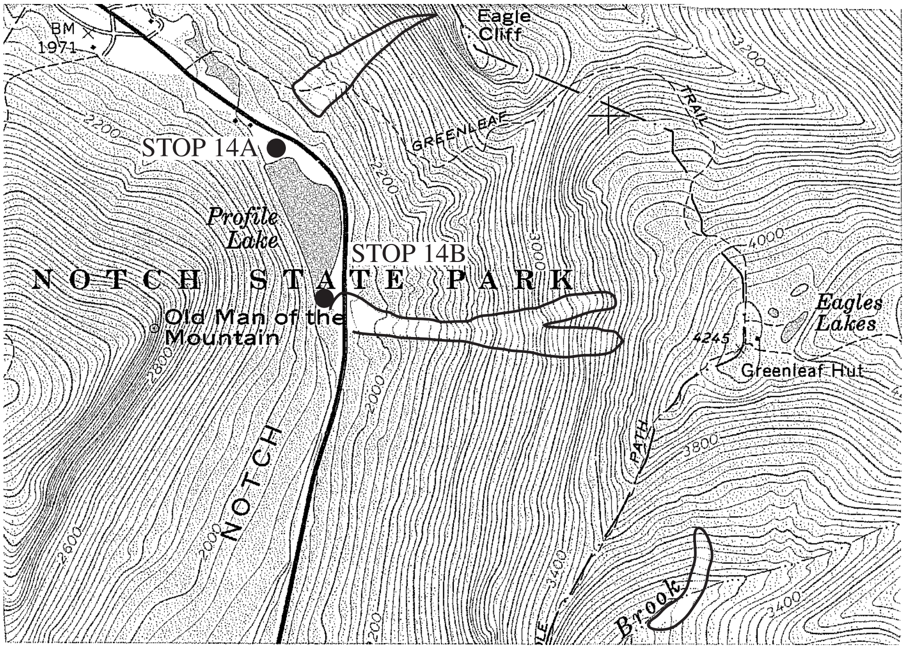

Figure 9: Part of the 7.5 minute Franconia quadrangle showing locations of stops and topographic setting of Profile Lake, Cannon Cliff, and landslide scars on Mt. Lafayette and Eagle cliff; contour interval is 40 feet.

These blocks are vertically separated from one another along sub-horizontal joints (N25°W, 23°NE) that are part of a single set of joints that form the top and bottom surfaces of all profile blocks. The Profile itself has been created by selective breakage at the edge of these blocks along nearly vertical joints (S65°W, 73°SE—the predominate set forming Cannon Cliff; N20°E, 75°SE; N35°E, 80°SE; N60°W, 60°SW; N25°E, Vertical; N20°W, 85°NE).

The information presented here is from a study conducted in 1976 as a part of the Environmental Impact Statement analyzing alternatives for Interstate-93 through Franconia Notch (Fowler, 1982). The specific intent of the study was to identify and assess possible damage to the Profile from the various construction alternatives then being considered, with particular attention to those related to blasting below on the floor of the Notch. The study concluded that limited blasting could be undertaken if closely monitored, and it drew the following conclusions about the stability of the Profile:

1. The principal mechanism responsible for the Profile’s stability is the compressive cantilevering created by the dead-weight load of most of the blocks comprising the Profile acting at the location of their combined center of gravity behind the cliff-face. 2. The blocks in the rock mass that are the most stable are those from the nose upward to the forehead, because their large intact masses and the location of the individual and combined centers of gravity well behind the cliff face act to enhance the cantilevering mechanism. 3. The blocks in the rock mass that are the least stable, and which pose the most important stability problems, are those from the upper-lip downward to the chin, because their comparatively small intact masses and the location of their individual centers of gravity very near to, or (in the case of the chin) in front of, the cliff-face are not part of the overall cantilevering mechanism and they obtain no support from the blocks that comprise the rest of the Profile. 4. Because of the positions and weights of the blocks in the Profile, the active mechanical reinforcement of the various portions of the rock mass to increase their security appears possible, but additional structural and mechanical analyses are needed to design a system that can be installed without endangering the Profile in the process.

These conclusions suggest that this additional work should be undertaken without delay. However, funds for such a project have not materialized, although interest in preserving the Profile has been keen since it was first discovered. So, for the foreseeable future, we can only hope that the natural mechanism(s) of stability within the rock mass itself will maintain the Profile’s security.

Efforts to secure the Profile with tie-rods and turnbuckles on the top of the forehead and with epoxy seals to keep water out of its structure have been philosophically reassuring but largely ineffective in improving the fundamental security of the Profile. But, there has been no lack of interest or effort in these attempts, and a summary of these remarkable activities over the last 190 years is presented in Fowler (1997). Readers looking for more information concerning the history of these maintenance efforts should consult Hancock (1980).

STOP 14B. Mass wasting processes (landslides and rockfalls), Franconia Notch. (Franconia quadrangle) Walk south from the Old Man viewpoint to the southern end of Profile Lake and a view of the spectacular slopes of Franconia Notch to the south (Fig. 9). The broad expanse of Cannon cliff dominates the western slope, and the bulky shoulder of Mt. Lafayette looms over the eastern slope of the notch. Mt. Lafayette is composed of Littleton schist. Cannon cliff is composed of the Conway granite, which exhibits textbook examples of exfoliation slabs, especially striking at the north end. A deep recess at the south end of the cliff owes its origin to a thick diabase dike, which trends roughly NNE through the higher notch separating Eagle cliff and Mt. Lafayette on the eastern slope. Cannon cliff is one of East's premier destinations for rock and ice climbing.

The south end of Profile Lake is dammed by landslide debris delivered from the west-facing slope of Mt. Lafayette in the area of the prominent landslide scar (Fig. 9). This form of mass wasting has been a very important mechanism in the modification of the landscape in Franconia Notch and other notches in the White Mountains during and following departure of late Wisconsinan ice.

As the regional climate rapidly warmed when continental ice was receding to the north, the newly exposed bedrock surfaces of the mountains and valleys were exposed to the harshly variable climate that existed close to the retreating glacier. This climate was characterized by large and frequent differences in diurnal temperatures, which permitted intense freezing and thawing of water in cracks and crevices in the newly exposed bedrock surfaces. This alternating freezing and thawing shattered the bedrock surfaces and created large volumes of blocky talus on the mountain peaks and slopes. As this talus continued to accumulate, it either slid or fell to the bases of the cliffs and valley walls, forming talus slopes particularly ubiquitous in Franconia Notch. This process lessened in intensity as the local climate continued to warm and stabilize, so the amount of shattering and rockfall has decreased but not stopped altogether.

The large talus slope at the base of the Cannon cliff (Fig. 9) is an excellent example of rockfall debris accumulation that took place as result of the freeze-thaw process, and although rockfall is not as intense today as it was 10,000 to 11,500 years ago, it is still continuing and poses a considerable safety threat to rock climbers on the cliff and even scrambling up to the base of the cliff. The now sharply pointed peaks of Franconia ridge (from north to south, Mts. Lafayette, Lincoln, Little Haystack, Liberty, and Flume) are the result of their glacially rounded peaks being shattered and sharpened by the freeze-thaw process. As the rock on the peaks was shattered and accumulated on mountain slopes, it slid off the higher parts of the peaks and accumulated on the lower flanks, creating peaked forms from the originally more-rounded summits.

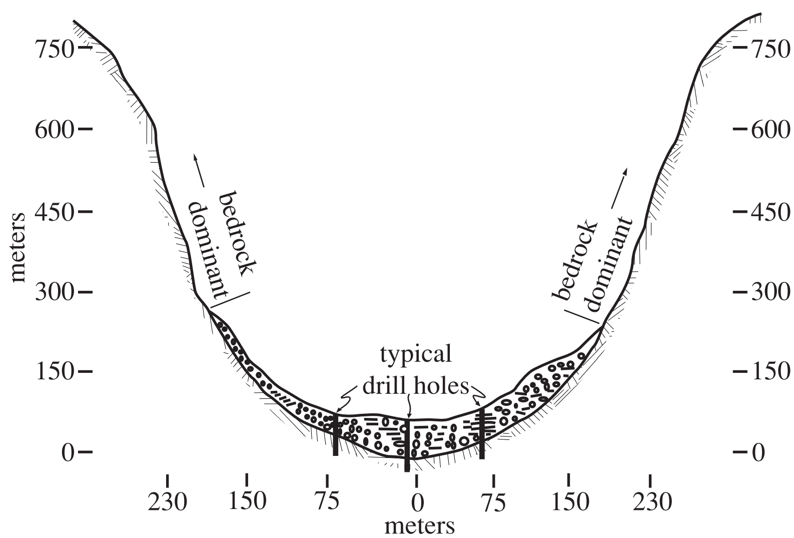

Figure 11: typical cross-section of the deepest part of Franconia Notch near the south end of Profile Lake.

These deposits of talus and other debris have accumulated since the ice sheet left the Notch and have in the process changed the topography of the glacial trough. Figure 10 is a typical cross-section of the deepest part of Franconia Notch near the south end of Profile Lake. The information in Fig. 10 was obtained from borings conducted for the pre-design engineering studies for I-93, and demonstrates how the glacially smoothed bedrock floor has been buried by mass wasting debris, leaving the rugged topography as seen today on the floor of the Notch.

Most all of the material on the floor of Franconia Notch is composed of, or derived form rocky debris from mass wasting, and many of the alternative construction designs for I-93 were abandoned because of the expense and difficulty associated with construction in such unconsolidated materials (e.g. cut-and-cover tunnel). Finally in the early 1980s, I-93 was completed as a parkway, and is the only non-four-lane section of the interstate highway system in the United States.

The process of debris creation high on the slopes and cliffs and their sliding or falling into the Notch are ongoing processes here and elsewhere around the White Mountains (Fowler, 1984; Waitt & Davis, 1988), and the many fresh landslide scars one can observe on the slopes of the Notch document their influence on the landscape. Highway crews frequently clean up landslide debris from the bottom of the many slide tracks that scar all three notches visited on this trip. However, Franconia Notch has been particularly problematic in this regard. Numerous historic landslides have occurred on Eagle cliff and the western flank of Mt. Lafayette (Flaccus, 1958), with major events occurring in 1826, 1850, 1883, 1915, 1947, 1948, 1959, and 1974. The landslides of 1948 and 1959 covered old US 3 with so much debris at this location that many days were required to re-open the highway. Many of these landslides were caused by extreme rainfall events, although the 1959 slide occurred during a particularly dry period in the autumn and may have been caused by highway improvements that removed too much toe from the earlier slides.

During the winters of 2000 and 2001, a research project was initiated to recover sediment cores from Profile Lake in order to reconstruct a pre-historic record of large landslides in Franconia Notch. A radiocarbon age from near the base of the longest sediment core from the middle of the lake suggests that sediment accumulation began over 10,000 years ago. From near the tops of sediment cores at the southern end of Profile Lake, layers of coarse debris that fine upwards thicken towards the eastern shore of the lake. These coarse layers probably reflect one or more historic landslides, and certainly the 1959 landslide that covered the highway and the mouth of the lake. All sediment cores are being analyzed at high resolution using magnetic susceptibility, loss on ignition, and grain size analyses to distinguish landslide-derived materials. Core tops will be analyzed for 210Pb in an attempt to distinguish historic events. Radiocarbon dating will be used to distinguish pre-historic events. Preliminary work suggests that coarse layers in sediment cores from Profile Lake, hence landslides, were more common during the early and late Holocene, and less frequent during the middle Holocene (Rogers et al., 2001). Assuming that most landslides are caused by extreme rainfall events (Sharpe, 1938), the sediment record from Profile Lake supports the work of Bierman et al. (1997), whose study of alluvial fans and pond sediments suggested that hillslope erosion in northern Vermont was caused by more severe storms during the early and late Holocene.

Cannon Mountain has one of the highest relief cliffs in the northeastern United States (~300 m). At about 10 a.m. on 19 June 1997, a large rockfall originated from near the top of the cliff's north end (GPS: 44°09'33.6"N; 71°41'05.8"W; elev. ~ 975 m). The release involved a large section of steeply inclined exfoliation slabs of Conway granite, which became airborne over the lower half of the cliff. The rockfall was apparently not triggered by seismic, sonic, or meteorological events, as observers did not note an earthquake, sonic boom, or quarry blast, and the previous four days were precipitation-free. Both the cliff scar and the impact swath on the talus slope below the cliff remain clearly visible from the valley floor.

Cannon cliff's talus slope, with a relief of about 250 m, is composed of angular blocks generally <5 m in length. Many parts of the talus slope are vegetated with trees and shrubs, but in open areas most talus blocks are covered with gray to black lichens, in contrast to the green lichens on the cliff face. Because of the altitude of the cliff, frost action is severe and rockfalls are common, especially during the spring.

The rockfall not only flattened all vegetation, including shrub birch and spruce, but mobilized most blocks for a width of about 100 m on the upper part of the talus slope, leaving immense exposures of fines (less than pebble size), which have begun to erode since June 1997. The talus slope absorbed most of the energy of impact, but many rockfall blocks traveled up to 425 m, cutting a 10-20 m wide, 5-10 m deep trough into the lower part of the slope. Beech and maple trees up to 45 cm diameter were splintered or sheared. One large rockfall block (5.1 m x 3.5 m x 3.3 m; ~3.5 tons) reached the valley floor, long axis oriented upslope, adjacent to a paved bicycle path ~380 m south of Profile Lake. This block originated from the cliff face, as part of the top surface remains mostly unscathed and covered with green lichens (Davis & Fowler, 1998).

The arrangement of blocks comprising the Old Man of the Mountain is susceptible to the same mass wasting processes demonstrated by the 1997 event, mechanisms that no doubt will be responsible for the Old Man's ultimate collapse, given its delicate state of stability.

Return to vehicles and continue south on the Franconia Notch Parkway.
62.1 note the spheroidal weathering of Conway Granite
64.1 Exit the parkway at sign for "The Basin." Follow marked walkways from the parking area.

STOP 15: The Basin, Franconia Notch. (Franconia quadrangle) Although not the largest and deepest in the White Mountains, the potholes at the Basin are the most accessible. We will conclude the surficial geological discussions on the field trip by addressing the questions about when and where potholes form. Conventional wisdom has suggested that potholes are created by the abrasive action of sand and gravel in swirling eddies of meltwater streams. However, cases may be made for an entirely subglacial origin of potholes, as they have been described on the top of a bedrock ridge in Ontario (Gilbert, 2001) that requires overlying ice to provide water flow in such a topographic setting. The same case may be made to explain potholes at an altitude of over 1200 meters along the Caps Ridge Trail on the western flank of Mt. Jefferson in northern Presidential Range in New Hampshire.

Much of the bedrock here is an intrusive breccia of the Jurassic Conway Granite intruding the early Devonian Kinsman Granodiorite. A sharper contact is exposed on the west side of the parkway, in the roadcut for the entrance ramp back on to Franconia Notch Parkway from the Basin parking lot on the northbound side.

Return to the vehicles and continue south on the Franconia Notch Parkway to I-93 and on to Boston.

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