Géographie physique et Quaternaire

Cirques of the Presidential Range, New Hampshire, and surrounding alpine areas in the northeastern United States

Les cirques du Presidential Range (New Hampshire) et des régions alpines avoisinantes du nord-est des États-Unis.

Alpine Vergletscherung und Kar-Morphometrie in der Presidential Range, New Hampshire, und in den umgebenden bergigen Gebieten in Nordost-U.S.A

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Résumé de l'article

Les signes de rajeunissement des glaciers de cirque après le retrait de l'inlandsis demeurent ambigus au Presidential Range (New Hampshire), au mont Katahdin et dans les Longfellow Mountains (Maine), ainsi dans les Adirondack (New York). Au Ritterbush Pond, dans le nord des Green Mountains du Vermont, les datations au radiocarbone de sédiments lacustres ont établi que l'extrémité d'une vallée de basse altitude, située en amont d'une série de moraines transversales, était libre de glace vers 11 940 BP. Bien que certains chercheurs croient que ces moraines témoignent d'une glaciation de cirque, il est aussi possible que les moraines aient été mises en place par une langue glaciaire pendant la déglaciation. Dans la vallée du Johnson Hollow Brook dans les Catskill (New York), une datation au radiocarbone de sédiments de base dans un étang fermé par une moraine indique que la glace pourrait avoir persisté jusqu'à vers 10 860 BP (Lederer et Rodbell, 1998). Comme cette moraine semble avoir été mise en place par un glacier de cirque, la datation fournit le meilleur indice de glaciation de cirque survenue après le retrait de l'inlandsis dans le nord-est des États-Unis. Les données sur la morphométrie des cirques, compilées à partir des nouvelles cartes topographiques, creusent le mystère en établissant que les deux sites des Green Mountains et des Catskill seraient moins propices au maintien des glaciers locaux après le retrait de l'inlandsis que le seraient d'autres cirques mieux calibrés et à plus haute altitude du Presidential Range et au mont Katahdin où on ne trouve pas d'indices de cirques glaciaires après le retrait de l'inlandsis.

Géographie physique et Quaternaire, 1999, vol. 53, n° 1, p. 25-45, 11 fig., 1 tabl., 1 app.

CIRQUES OF THE PRESIDENTIAL RANGE, NEW HAMPSHIRE, AND SURROUNDING ALPINE AREAS IN THE NORTHEASTERN UNITED STATES

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ABSTRACT Evidence for rejuvenation of cirque glaciers following wastage of continental ice remains elusive for the Presidential Range and Mount Moosilauke of New Hampshire, Mount Katahdin and the Longfellow Mountains of Maine, and the Adirondack Mountains of New York. At Ritterbush Pond in the northern Green Mountains of Vermont, radiocarbon ages from lake sediment cores suggest that a low-altitude valley head, located up-valley of a series of cross-valley moraines, was ice-free by 11,940 ^{14}C yrs BP (Bierman *et al.*, 1997). Although some workers argue that these moraines in Vermont are evidence for cirque glaciation, the moraines could have been formed by a tongue of continental ice during deglaciation. At Johnson Hollow Brook valley in the Catskill Mountains of New York, a radiocarbon age from basal sediments in a pond dammed by a moraine suggests that glacier ice may have persisted until 10,860 ^{14}C yrs BP (Lederer and Rodbell, 1998). Because this moraine appears to have been deposited by a cirque glacier, the radiocarbon age provides the best evidence in the northeastern United States for cirque glaciation post-dating recession of continental ice. Cirque morphometric data, compiled from newly available topographic maps, add to the conundrum that these two sites in the Green and Catskill Mountains should not be nearly as favorable for maintaining local glaciers post-dating icesheet recession as higher-altitude and better-developed cirques in the Presidential Range and Mount Katahdin, where evidence for post-icesheet cirque glaciers is lacking.

RÉSUMÉ *Les cirques du Presidential Range (New Hampshire) et des régions alpines avoisinantes du nord-est des États-Unis.* Les signes de rajeunissement des glaciers de cirque après le retrait de l'inlandsis demeurent ambigus au Presidential Range (New Hampshire), au mont Katahdin et dans les Longfellow Mountains (Maine), ainsi dans les Adirondack (New York). Au Ritterbush Pond, dans le nord des Green Mountains du Vermont, les datations au radiocarbone de sédiments lacustres ont établi que l'extrémité d'une vallée de basse altitude, située en amont d'une série de moraines transversales, était libre de glace vers 11 940 BP. Bien que certains chercheurs croient que ces moraines témoignent d'une glaciation de cirque, il est aussi possible que les moraines aient été mises en place par une langue glaciaire pendant la déglaciation. Dans la vallée du Johnson Hollow Brook dans les Catskill (New York), une datation au radiocarbone de sédiments de base dans un étang fermé par une moraine indique que la glace pourrait avoir persisté jusqu'à vers 10 860 BP (Lederer et Rodbell, 1998). Comme cette moraine semble avoir été mise en place par un glacier de cirque, la datation fournit le meilleur indice de glaciation de cirque survenue après le retrait de l'inlandsis dans le nord-est des États-Unis. Les données sur la morphométrie des cirques, compilées à partir des nouvelles cartes topographiques, creusent le mystère en établissant que les deux sites des Green Mountains et des Catskill seraient moins propices au maintien des glaciers locaux après le retrait de l'inlandsis que le seraient d'autres cirques mieux calibrés et à plus haute altitude du Presidential Range et au mont Katahdin où on ne trouve pas d'indices de cirques glaciaires après le retrait de l'inlandsis.

ZUSAMMENFASSUNG *Alpine Vergletscherung und Kar-Morphometrie in der Presidential Range, New Hampshire, und in den umgebenden bergigen Gebieten in Nordost-U.S.A.* Hinweise auf eine Verjünjung der Kar-Gletscher als Folge des Schwunds von kontinentalem Eis bleiben für die Presidential Range von New Hampshire, den Mount Katahdin und die Longfellow Mountains von Maine sowie die Adirondacks von New York schwer greifbar. Am Ritterbush Pond in den nördlichen Green Mountains von Vermont zeigen Radiokarbondatierungen von Seesedimentkernen, dass ein Talhaupt von niedriger Höhe, das sich oberhalb einer Serie von Quermoränen befand, um etwa 11 940 ^{14}C Jahre v.u.Z. (Bierman *et al.*, 1997) eisfrei war. Obwohl einige Forscher diese Moränen für einen Beweis der Kar-Vergletscherung halten, könnten diese Moränen durch eine Zunge kontinentalen Eises während der Enteisung gebildet worden sein. Im Johnson Hollow Brook-Tal in den Catskill Mountains von New York erlaubt eine Radiokarbondatierung von Basis-Sedimenten in einem von einer Moräne eingedämmten Teich die Annahme, dass Gletscher-Eis bis etwa um 10 860 ^{14}C Jahre v.u.Z. (Lederer und Rodbell, 1998) gedauert hat. Da diese Moräne durch einen Kar-Gletscher abgelagert worden zu sein scheint, liefert die Radiokarbondatierung den besten Beweis für Kar-Vergletscherung nach dem Rückzug des Kontinental-Eises im Nordosten der U.S.A. Morphometrische Kar-Daten, die mittels neu verfügbarer topographischer Karten zusammengestellt wurden, tragen zu dem Rätsel bei, dass nämlich diese zwei Plätze in den Green und Catskill Mountains für die Aufrechterhaltung lokaler Gletscher nach dem Rückzug der Eisdecke bei weitem nicht so geeignet hättensein sollen wie höher gelegene und besser entwickelte Kare in der Presidential Range und am Mount Katahdin, wo Belege für Kar-Gletscher nach dem Rückzug der Eisdecke fehlen.

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INTRODUCTION

The purpose of this report is twofold: 1) to evaluate evidence for local (cirque) glaciation in the northeastern United States (Fig. 1) in light of work since the publication by Waitt and Davis (1988), and 2) to summarize data on cirque morphometry, especially schrund altitudes (Fig. 2), measured on new topographic base maps that have become available in the past 10 years for the Presidential Range in New Hampshire, the Longfellow Mountains and Mount Katahdin in Maine, selected sites in the Green Mountains of Vermont, as well as the Adirondack and Catskill Mountains in New York. Although the base map has been available for over 30 years, complete cirque morphometric data from Mount Moosilauke in New Hampshire are summarized for the first time. Many of the ideas in this paper have been presented during the past decade on society field trips in the Presidential Range (Davis *et al.*, 1988, 1993, 1996a), Mount Katahdin (Davis and Caldwell, 1994), and the Green Mountains (Wright *et al.*, 1997b). Cirque morphometry maps are included in this paper; photographs and geologic maps are not included, although reference is made to the literature where appropriate.

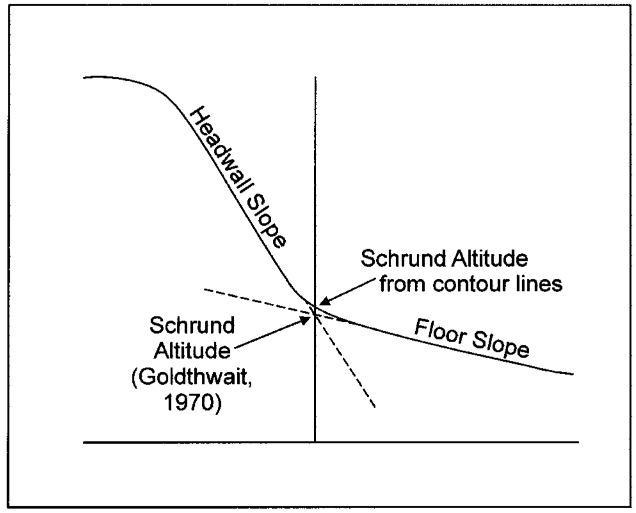


FIGURE 2. Schematic illustration for estimation of schrund altitude (after Goldthwait, 1970, Fig. 9, p.100).

Estimation schématisée de l'altitude d'une rimaye (selon Goldthwait, 1970, fig. 9, p. 100).

RELATIVE AND RADIOMETRIC CIRQUE GLACIER CHRONOLOGIES

PRESIDENTIAL RANGE, NEW HAMPSHIRE

The name Goldthwait is synonymous with not only the glacial history of New Hampshire (Goldthwait *et al.*, 1951) but also with our understanding of cirque glaciation in the Presidential Range (Figs. 1, 3) of the White Mountains (see W.B. Thompson, this volume). James W. Goldthwait was the first to carry out an extensive study of glaciation in the Presidential Range (1913a, b, 1916, 1938), where he reached three major conclusions: 1) the uplands above the cirques were eroded by both fluvial and glacial activity, 2) the cirques were carved by alpine glaciers, as opposed to continental ice, stream erosion, or frost action, and 3) continental glaciation followed the last cirque glacier activity. J.W. Goldthwait's 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. J.W. Goldthwait (1913a) did not support the concept that local glaciers extended far down valleys from an icecap centered on the Presidential Range, as proposed by Packard, Vose, and the Hitchcocks (see W.B. Thompson, this volume).

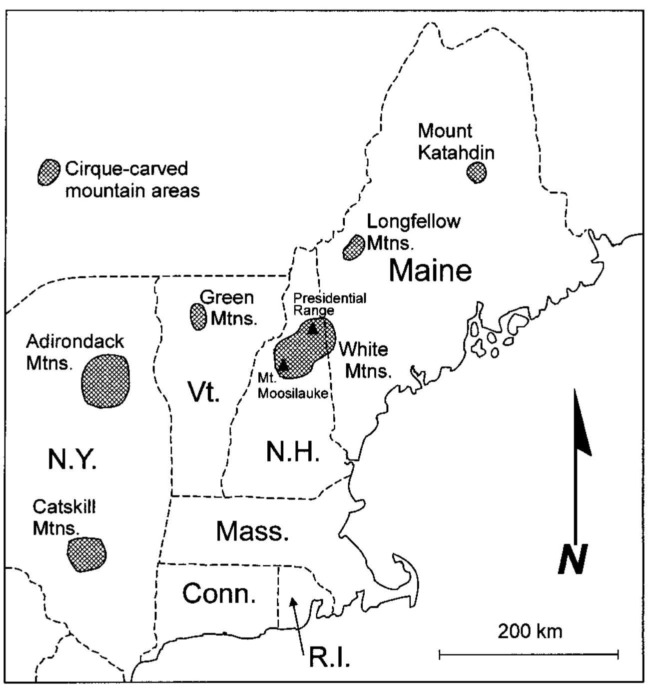


FIGURE 1. Index map for alpine areas with cirques in northeastern United States (after Bradley, 1981, and Ackerly, 1989).

Carte repère du nord-est des États-Unis illustrant les régions alpines où se trouvent des cirques (selon Bradley, 1981 ; Ackerly, 1989).

Over the next two decades, only two workers strongly disputed the conclusions of J.W. Goldthwait concerning the timing of continental and cirque glaciation in the Presidential Range. Johnson (1917, 1933) suggested that 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 and on Mount Katahdin, Maine; however, neither author provided a convincing explanation for the till of farther northern provenance on the cirque floors in the two areas.

Richard P. Goldthwait (1939, 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's in support of cirque glacier

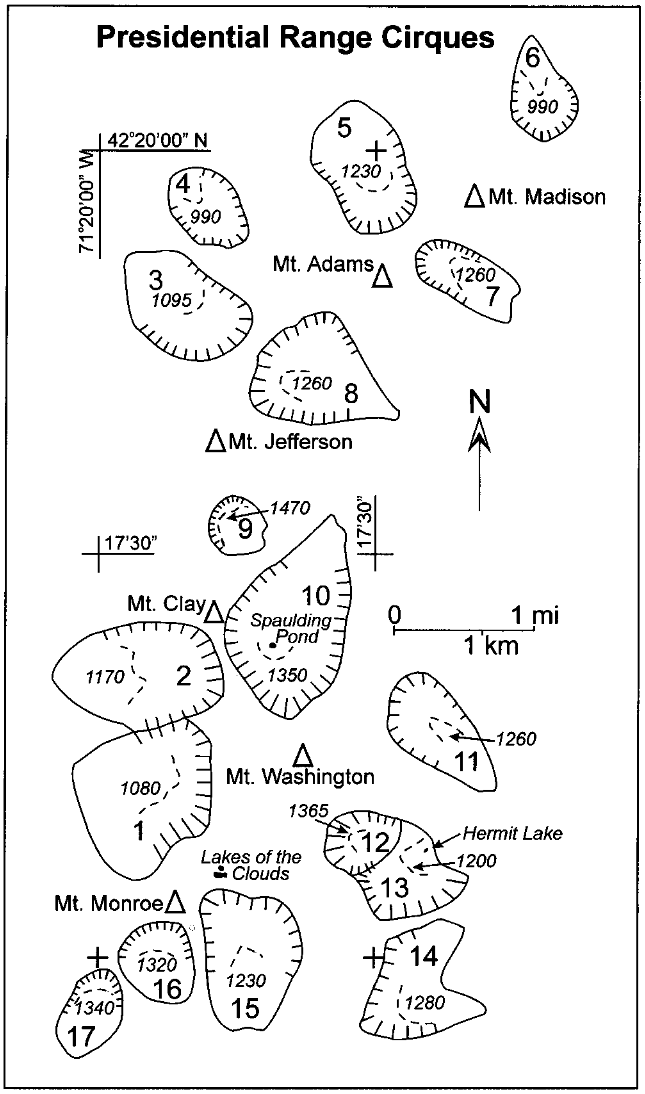


FIGURE 3. Cirque morphometric map for Presidential Range. Solid lines indicate outlines of cirques, hachured solid lines are cirque headwalls, dashed lines are cirque schrunds, with altitudes indicated in meters. Numbers refer to Table I. Small water bodies shown in black for cirques #10 and 13.

Morphométrie des cirques du Presidential Range. Les lignes pleines font le pourtour des cirques, les lignes hachurées montrent les murs de rimaye et les tirets représentent les rimayes (altitude en m). Les chiffres se rapportent au tableau I. Les cirques n^{os} 10 et 13 comprennent chacune une petite étendue d'eau.

activity preceding the last overriding by continental ice, but he also observed roche moutonnées on cirque floors along with striae and grooves on cirque headwalls, which he believed could only have been formed by continental ice. In his 1970 paper, R.P. Goldthwait reviewed his earlier work and provided pebble lithology data for till sites on the uplands and in north-facing cirques, which supported the view that till in the Presidential Range was deposited by continental ice. Also, in this latter paper, R.P. Goldthwait presented morphometric data on cirques and altitudinal estimates of firn lines for the former cirque glaciers in the Presidential Range. 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 1950s, W.F. Thompson (1960a, b, 1961) analyzed aerial photographs of the Presidential Range and Mount Katahdin in Maine and refuted the Goldthwaits' view by arguing that the steep headwalls and sharp arêtes were indicative of active cirque glaciers following continental icesheet deglaciation. W.F. Thompson's primary field contribution was an experiment to test the origin of striae whereby he painted one of the striated portions of the Tuckerman Ravine headwall. Although W.F. Thompson did not present field data to support his view, he believed that moraines of cirque glaciers had been obliterated by post-glacial mass wasting processes. Work in Tuckerman Ravine during the late 1980s by D.J. Thompson (1990, this volume) 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 to the south. However, Gerath and Fowler (1982), Fowler (1984), Gerath *et al.* (1985), Davis and Waitt (1986), and Waitt and Davis (1988) examined the diamicts at the cirque mouths and concluded that the sediments are not till, but rather debris flow deposits. Bradley (1981) also noted fresh grooves across the painted surface on the Tuckerman Ravine headwall, suggesting that if recent snow/ice avalanches could erode bedrock, then perhaps cirque glaciers could also striate cirque headwalls. In 1998, few cobbles and boulders remain on W.F. Thompson's painted surface, and paint is only preserved in the deepest grooves. However, paint also does not survive 30 years on trail signposts exposed to the severe weather conditions in the Presidential Range, so the significance of W.F. Thompson's experiment remains uncertain.

Opportunities for developing a radiocarbon chronology for the deglaciation of cirques in the Presidential Range are limited because of the small number of tarns. Spaulding Pond in the Great Gulf and Hermit Lake in Tuckerman Ravine (Fig. 3), although shallow, may provide useful continuous post-glacial records of sediment accumulation and should be cored. Lakes close by cirques in the Presidential Range have provided minimum radiocarbon ages for ice retreat (Davis *et al.*, 1980; Spear, 1989; Spear *et al.*, 1994; Miller and Spear, this volume).

Organic material from sediments at the base of a core retrieved from Lost Pond at an elevation of 650 m in Pinkham Notch on the east side of the Presidential Range (just off right margin of Fig. 3) provide a radiocarbon age of 12,870 ± 370 yrs BP (QL-985; Spear *et al.*, 1994; all ages reported in this paper are in ^{14}C yrs BP). 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 (Fig. 3) have 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 (Fig. 4), which provides a radiocarbon age of 13,000 ± 400 yrs BP (QL-1133; Spear, 1989). Given the model that continental ice thinned, separated, stagnated, and dissipated over the mountains of northern New England during Late Wisconsinan deglaciation (Goldthwait and Mickelson, 1982; Hughes *et al.*, 1985; Stone and Borns, 1986; Borns, 1987; Davis and Jacobson, 1987; Thompson and 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.

Current work by the author and Paul Bierman at the University of Vermont 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. Included are samples from two large boulders on the tongue-shaped rock glacier on the floor of Tuckerman Ravine to determine the relative age of this cirque deposit (see D.J. Thompson, this volume). Laboratory preparation and analyses of these samples are ongoing.

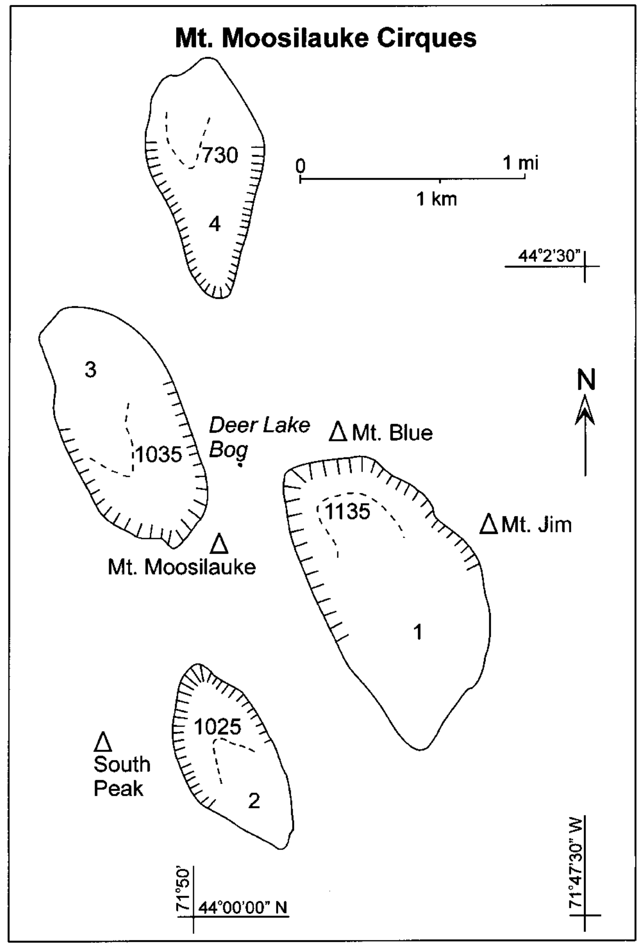


FIGURE 4. Cirque morphometric map for Mount Moosilauke. Solid lines indicate outlines of cirques, hachured solid lines are cirque headwalls, dashed lines are cirque schrunds, with altitudes indicated in meters. Numbers refer to Table I.

Morphométrie des cirques au mont Moosilauke. Les lignes pleines font le pourtour des cirques, les lignes hachurées montrent les murs de rimaye et les tirets représentent les rimayes (altitude en m). Les chiffres se rapportent au tableau I.

MOUNT MOOSILAUKE, NEW HAMPSHIRE

Mount Moosilauke is the second highest massif in the western White Mountains of New Hampshire (Figs. 1, 4). Of all other cirques in the White Mountains outside the Presidential Range, those on Mount Moosilauke are the best-developed. Haselton (1975) described evidence for continental ice overriding the summit areas and noted three cirques on Mount Moosilauke: Jobildunk, Gorge Brook, and Benton Ravines. Although he did not recognize any moraines on cirque floors, Haselton (1975) remained open to the possibility that cirque glaciers post-dated recession of continental ice. Unfortunately, none of the cirques on Mount Moosilauke have tarns that might yield sediment cores for radiocarbon dating.

MOUNT KATAHDIN, MAINE

Nearly all previous researchers at Mount Katahdin (Tarr, 1900; Antevs, 1932; Thompson, 1960a, b, 1961; Caldwell, 1966, 1972, 1998; Caldwell and Hanson, 1982, 1986) have promoted steep headwalls and sharp arêtes as evidence that cirque glaciers post-date continental icesheet deglaciation here (Figs 1, 5); they believe such features could not withstand the effects of an overriding icesheet. These workers also interpreted landforms on cirque floors as moraines and the moraine damming Basin Ponds on the east flank of Mount Katahdin (Fig. 5) to be formed, at least in part, by cirque glaciers. However, Davis (1976, 1978, 1983, 1989) reported observations from Mount Katahdin similar to those made in the Presidential Range by the Goldthwait's as evidence against post-icesheet cirque glacier activity. These observations included: 1) a lack of looped moraines on cirque floors, with the bumps in topography on cirque floors noted by others being hummocky till or landslide deposits rather than moraines, 2) till of a northern provenance on all cirque floors, with especially high percentages of erratic pebble lithologies on the floor of Northwest Basin, a northwest-facing cirque, and 3) roche moutonnées indicating upvalley ice flow on the floor of Northwest Basin. Although striae were not found on cirque headwalls, glacially lpolished surfaces were noted about halfway up Cathedral arête (Davis, 1976) and on Knife Edge arête (Davis *et al.*, 1996b). Along with

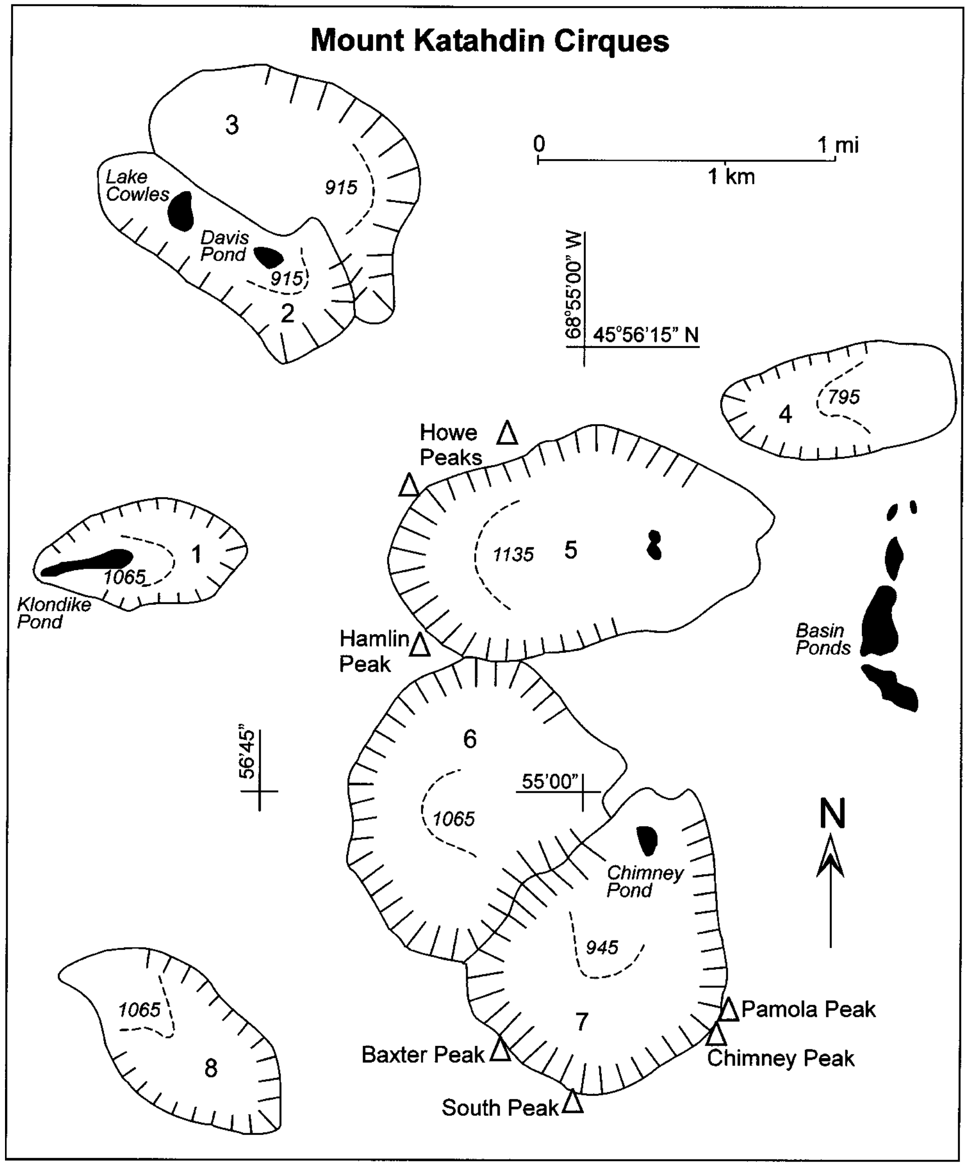


FIGURE 5. Cirque morphometric map for Mount Katahdin, Maine. Solid lines indicate outlines of cirques, hachured solid lines are cirque headwalls, dashed lines are cirque schrunds, with altitudes indicated in meters. Cirque numbers refer to Table I. Water bodies shown in black for cirques #1, 2, 5, and 7, and along Basin Ponds moraine east of cirques #5, 6, and 7.

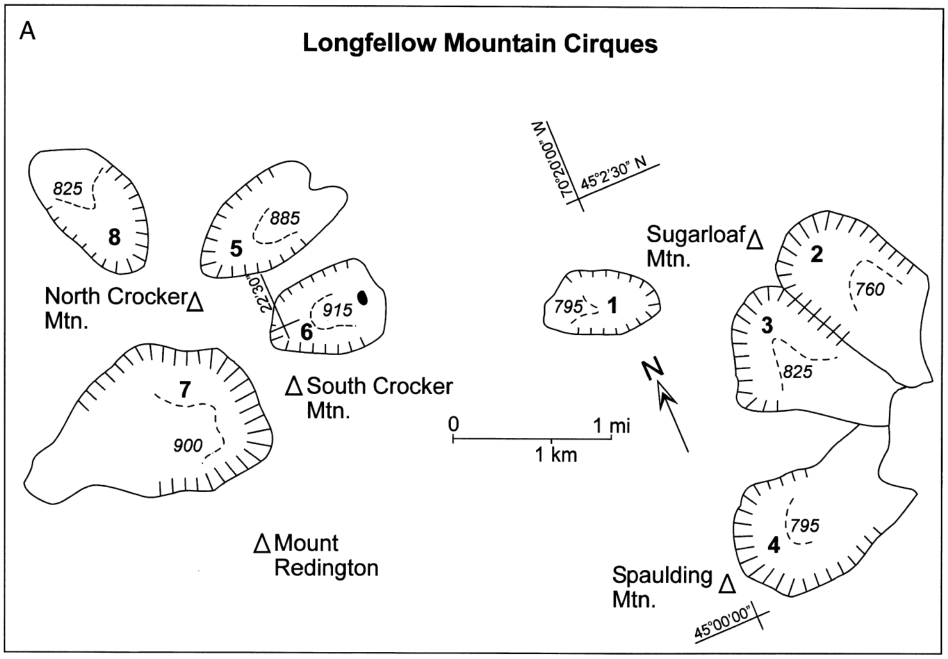
Morphométrie des cirques au mont Katahdin. Les lignes pleines font le pourtour des cirques, les lignes hachurées montrent les murs de rimaye et les tirets représentent les rimayes (altitude en m). La numérotation des cirques est celle du tableau I. Les petites étendues d'eau sont en noirs dans les cirques n^os^ 1, 2, 5 et 7 et le long de la moraine de Basin Pond à l'est des cirques n^os^ 5, 6 et 7.

thinness of soils, limited weathering of erratics located near the summit, and theoretical ice profiles, the glacially polished surfaces suggest that Mount Katahdin was overridden by a warm-based continental icesheet at some time during the late Wisconsinan glaciation (Davis, 1989).

Davis (1976, 1978, 1983, and 1989) argued that the Basin Ponds moraine was completely formed by continental ice to the east because: 1) the pebble fraction is 10 to 44% erratic, 2) its morphology is convex westward and follows a contour along the east slope of Mount Katahdin, 3) the moraine extends both north and south beyond the mouths of the three east-facing cirques, 4) there is too little space (< 20 m) between the moraine and Keep Ridge for a cirque glacier, where only a small drainage channel occurs, 5) several smaller arcuate and parallel ridges downslope of the Basin Ponds moraine that lie north and south of the cirque mouths could only be formed by receding continental ice to the east, and 6) along the south slope of Mount Katahdin, where there are no cirques, nearly continuous ridges extend for about 8 km at about the same altitude of the Basin Ponds moraine, so could only be formed by continental ice. Thus, the Basin Ponds moraine and the moraines on the south slope of Mount Katahdin were most likely formed during a late-glacial (nunatak) phase of late Wisconsinan glaciation.

Based on field data from cirques, Davis (1976, 1978) argued that the Late Wisconsinan regional snow-line on Mount Katahdin was too high to support formation of cirque glaciers following icesheet recession, the same case argued by R.P. Goldthwait (1970) for the Presidential Range.

No radiocarbon dating besides that presented in Davis and Davis (1980) is available for lakes and bogs in the Mount Katahdin area. However, preliminary cosmogenic nuclide analyses of ^{10}Be and ^{26}Al from one boulder on a recessional moraine (P. R. Bierman, oral commun., 1998) suggest that the earliest radiocarbon ages from bog and pond basal sediments on the moun-



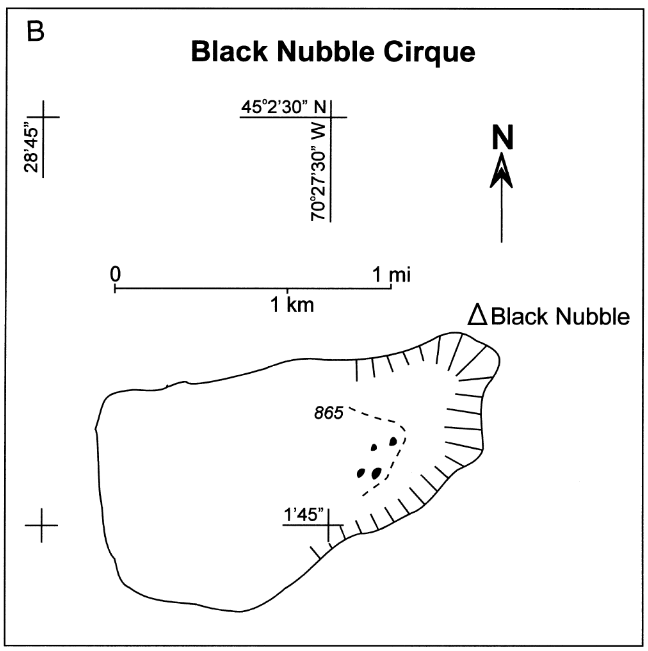


FIGURE 6. A. Cirque morphometric maps for Longfellow Mountains, Maine. Solid lines indicate outlines of cirques, hachured solid lines are cirque headwalls, dashed lines are cirque schrunds, with altitudes indicated in meters. Numbers for cirques on Sugarloaf and Crocker Mountains refer to Table I; only cirque #6 has small water body, shown in black. B. Black Nubble cirque; water bodies shown in black.

A. Morphométrie des cirques des Longfellow Mountains, dans le Maine. Les lignes pleines font le pourtour des cirques, les lignes hachurées montrent les murs de rimaye et les tirets représentent les rimayes (altitude en m). La numérotation des cirques est celle du tableau I ; seul le cirque n^{o} 6 renferme une petite étendue d'eau (en noir). B. Le cirque Black Nubble ; les étendues d'eau sont en noir.

tain lag ice retreat by several thousand years, as suggested by Davis and Davis (1980). Additional samples of polished bedrock from the Knife Edge arête (Davis *et al*, 1996b), along with boulders on cirque floors, lateral moraines, and the lowlands surrounding Mount Katahdin, are being analyzed for ^{10}Be and ^{26}Al.

LONGFELLOW MOUNTAINS, MAINE

In their study of glaciation in west-central Maine, Borns and Calkin (1977) concluded that ten cirque-like basins in the Longfellow Mountains (Figs. 1, 6) showed no evidence for reactivation by local ice during or subsequent to dissipation of Late Wisconsinan continental ice. Deposits and landforms related to these basins could be explained by continental glaciation and subsequent stagnation of this ice. No looped moraines or similar deposits were found in these cirques. Furthermore, examination of clast content in till forming the floor of two of the best-formed cirques, these facing eastward on Crocker Mountain (Fig. 6a), revealed about 50% erratic lithologies. Based on such data, Borns and Calkin (1977) concluded that the regional snow-line rose to an altitude above the Longfellow Mountains prior or during their emergence from the receding continental icesheet.

Borns and Calkin (1977) reported radiocarbon ages from the lowlands in the Longfellow Mountains between 10,030 ± 180 yrs BP (CY-2464) and 10,860 ± 160 yrs BP

(M. Stuiver, written commun., 1970). These ages are supported by more recently obtained radiocarbon ages for basal sediments from Boundary Pond (11,200 ± 200 yrs BP, GSC-1248, Shilts, 1981) and from Lower Black Pond (11,500 ± 50 yrs BP, OS-7123, Thompson *et al.*, 1996). The latter is an AMS radiocarbon age on terrestrial macrofossils; the former is a conventional radiocarbon age on bulk sediment. Small ponds in two cirques on Crocker Mountain (Fig. 6a) have not been investigated for sediment coring potential. All of these radiocarbon ages are consistent with recent chronological work on glacial Lake Hitchcock (Ridge and Larsen, 1990; Ridge *et al.*, 1996), which led to a new deglaciation model for northern New England proposed by Ridge *et al.* (this volume), who suggest that continental ice remained up to 1500 ^{14}C yrs later than suggested by existing chronologies. However, other radiocarbon ages for deglaciation of the White Mountains support a model for earlier recession of continental ice (Thompson *et al.*, this volume).

GREEN MOUNTAINS, VERMONT

Stewart (1961, 1971) and Stewart and MacClintock (1969) interpreted all drift in the Green Mountains of northern Vermont as the product of a continental icesheet. However, Wagner (1970, 1971) and Connally (1971) proposed that local mountain glaciers post-dated icesheet recession (Figs. 1, 7, 8). These latter authors argued that some low-elevation valley heads are cirques, some lakes are tarns, some ridges on valley floors are moraines built by cirque glaciers, and some deltas down-valley of the ridges were formed by meltwater of cirque glaciers. Waitt and Davis (1988) questioned the conclusions of Wagner (1970, 1971) and Connally (1971), noting that the valley heads do not have the typical bowl shape of cirques, the lakes are too large to be tarns, some of the ridges do not loop across basin floors typical of end moraines formed by cirque glaciers, the deltas are similar to many other ice-contact deposits in the area unrelated to valley heads and therefore are better explained by meltwater from continental ice, and the regional snow-line was too high during the Late Wisconsinan glaciation to support local glaciers at the valley heads. Waitt and Davis (1988) did identify five to seven high-elevation cirques on Mount Mansfield that are comparable to cirques elsewhere in the northeastern United States. Thus, Waitt and Davis (1988) concluded that all of the glacial landforms described by Wagner (1970, 1971) and Connally (1971) were more readily explained by tongues of continental ice rather than by cirque glaciers.

Stephen Wright and his students from the University of Vermont (Wright *et al.*,1997a, b; Loso et al, 1998) re-examined the ridges in the Miller Brook valley (Fig. 7) and concluded that upper valley ridges that were interpreted to be moraines by Wagner (1970) are part of an esker that extends the full length of the valley, ending in a subaqueous fan deposit (Wagner's delta). Wright *et al.* (1997a, b) and Loso *et al.* (1998) also suggested that summer temperatures must drop by about 14°C to support a cirque glacier in Miller Brook valley.

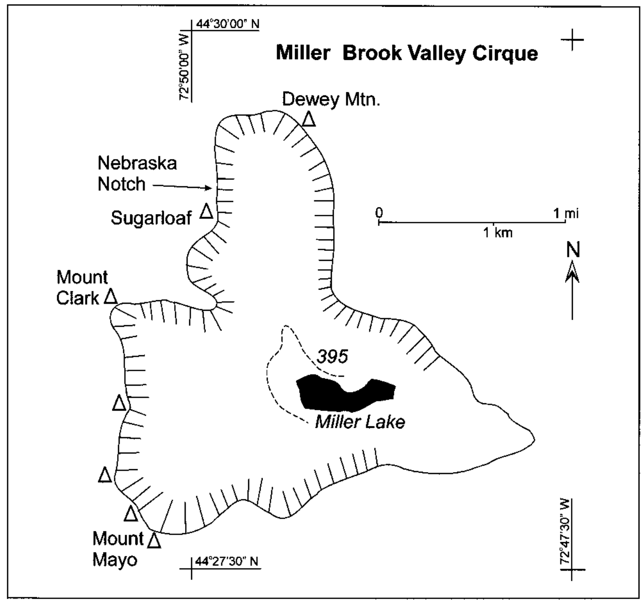


FIGURE 7. Cirque morphometric map for Miller Brook valley cirque, Green Mountains, Vermont. Solid line indicates outline of cirque, hachured solid line is cirque headwall, dashed line is cirque schrund, with altitude indicated in meters. Miller Lake is labeled.

Morphométrie du cirque de la vallée du Miller Brook, dans les Green Mountains (Vermont). Les lignes pleines font le pourtour du cirque, les lignes hachurées montrent les murs de rimaye et les tirets représentent les rimayes (altitude en m).

Sperling *et al.* (1989) interpreted radiocarbon ages obtained from sediment cores in an ephemeral pond alongside the moraine-like feature (esker) in the Miller Brook valley to support the conclusions of Wagner (1970, 1971) and Connally (1971) that cirque glaciers post-dated icesheet deglaciation in the northern Green Mountains. They radiocarbon dated the 275-285 cm interval of one of the sediment cores in Miller Brook valley at 9,280 ± 235 yrs BP (QC1273A), which is considerably younger than other bog- and pond-bottom radiocarbon ages in northern New England. However, Sperling *et al.* (1989) made no mention of the material that they dated or whether gray inorganic silt typical of the late-glacial parts of lake sediment cores was recovered, and thus it is likely that this radiocarbon age underestimates the time of deglaciation.

Wagner (1970) also mapped a series of cross-valley moraines in Ritterbush valley and suggested that the upper part of the valley was a site for cirque glaciation following recession of continental ice (Fig. 8). Sperling *et al.* (1989) recovered lake sediments from a core through a post-glacial delta at the west end of Ritterbush Pond. They radiocarbon-dated the 840-850 cm interval at 10,730 ± 200 yrs BP (QC1272A) and the 850-860 cm interval at 10,090 ± 230 yrs BP (QC1272B). Sperling *et al.* (1989) interpreted the older age to be the more accurate for deglaciation of the valley and suggested that cirque glaciers were present in the valley until about 11,000 yrs ago. However, their pollen analysis of the radiocarbon-dated basal sediments indicates that tundra and spruce-fir pollen zones typical of basal zones of most pollen diagrams in the New England area are not present, suggesting that hundreds if not thousands of years of the post-glacial sediment record may be missing. Again, Sperling *et al.* (1989) did not note whether gray inorganic silt was recovered from the basal part of their Ritterbush Pond sediment core.

In an effort to clarify the timing of ice retreat from the Green Mountains, the author obtained two overlapping sediment cores from the middle of Ritterbush Pond (317 m), with Paul Bierman, Andrea Lini, and their students at the University of Vermont (Lini *et al.*, 1995; Lin *et al.*, 1995; Bierman *et al.*, 1997). Bulk sediments from 569 cm below the mud-water interface (in Ritterbush Core 2) were AMS radiocarbon-dated at 21,860 ± 370 yrs BP (CAMS 20197, $\delta^{13}C = -24‰$). The 479 cm depth of this core was AMS radiocarbon-dated 11,940 ± 90 yrs BP (CAMS 20902; corrected for $\delta^{13}C = -34‰$). Lini *et al.* (1995) argue that the $\delta^{13}C$ values of total organic carbon do not indicate terrestrial vegetation as a major component of primary productivity until about 12,000 ^{14}C yrs ago, thus the 21,860 yrs BP radiocarbon age could reflect a different source of older carbon than the younger ages. From the basal parts of lake sediment cores elsewhere in New England, similarly old radiocarbon ages to the 21,860 ± 370 yrs BP age from Ritterbush Pond have been noted (Davis and Davis, 1980; Davis *et al.*, 1995b; see Ridge *et al.*, this volume). In all cases, these old radiocarbon ages are from sediments with very low values of total organic carbon as determined by loss-on-ignition analyses (Lini *et al.*, 1995; Bierman *et al.*, 1997; P.T. Davis, unpublished data). Five high-altitude cirques on Mount Mansfield (Fig. 9) do not include tarns or other suitable sites for sediment coring and radiocarbon dating.

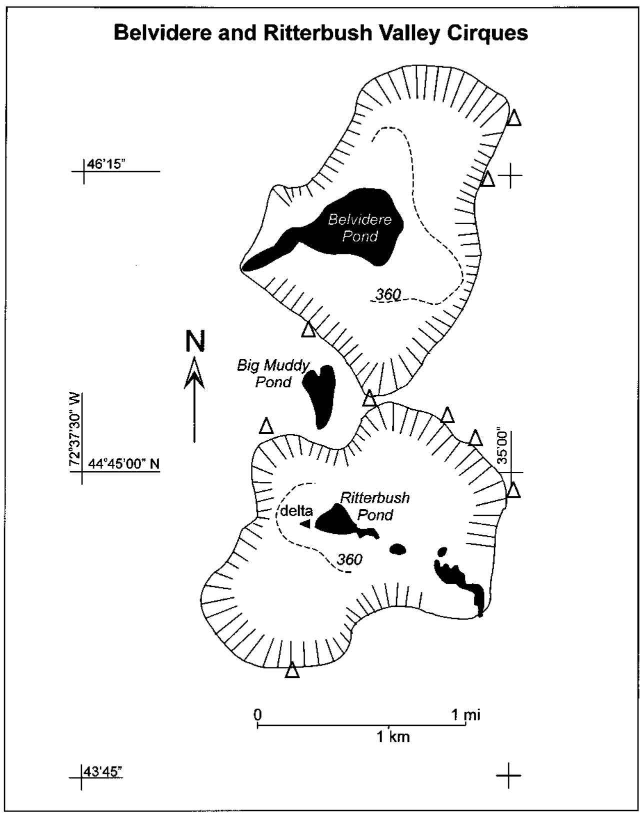


FIGURE 8. Cirque morphometric map for Belvidere (north) and Ritterbush valley (south) cirques, Green Mountains, Vermont. Solid lines indicate outlines of cirques, hachured solid lines are cirque headwalls, dashed lines are cirque schrunds, with altitudes indicated in meters. Water bodies shown in black, except for Belvidere and Ritterbush ponds, which are labeled.

Morphométrie des cirques Belvidere (au nord) et de la vallée du Ritterbush (au sud), dans les Green Mountains (Vermont). Les lignes pleines font le pourtour des cirques, les lignes hachurées montrent les murs de rimaye et les tirets représentent les rimayes (altitude en m). Les étendues d'eau sont en noir.

We also obtained sediment cores from Sterling Pond (917 m), about 60 km south of Ritterbush Pond, in order to compare pollen records from sites at different altitudes (Lin *et al.*, 1995). Bulk sediments from the 522 cm depth near the base of one sediment core from Sterling Pond were radiocarbon-dated at 12,760 ± 70 yrs BP (CAMS 17895). Taken at face value, the radiocarbon ages from Ritterbush and Sterling Ponds in the northern Green Mountains are remarkably similar to those from comparable altitudes in the White Mountains and suggest that continental ice thinned and/or back-wasted rapidly during Late Wisconsinan deglaciation. Palynological analyses from Ritterbush and Sterling Ponds are in progress to determine whether pollen indicators might be useful as a chronological tool where radiocarbon ages are suspect (Lin *et al.*, 1995; Davis *et al.*, unpublished data).

ADIRONDACK MOUNTAINS, NEW YORK

Craft (1976, 1979) examined numerous sites in the High Peaks region of the Adirondack Mountains (Fig. 1) that he suggested were locations of cirque glacier activity during and following disintegration of continental ice. Craft's evidence included over-deepened basins, steep headwalls and side walls, and moraines on basin floors. However, Barclay (1993) re-examined three of the main sites described by Craft (1976) and concluded that the evidence for local glaciers post-dating retreat of continental ice was weak. The Lost Pond depression adjacent to Weston Mountain was found to be lacking key morphometric elements of a cirque, and mapping of glacial lake sediments in the East Roaring Brook valley on the east side of Giant Mountain placed severe constraints on the extent of any local glacier in this valley, assuming that any post-Wisconsinan local glacier existed here at all (Barclay, 1993).

White Brook valley on the northeast side of Whiteface Mountain (Fig. 10) has long been considered to have hosted a local glacier following wastage of continental ice (Alling, 1916, 1919; Johnson, 1917; Craft, 1976, 1979). However, Barclay (1993) showed that an unvegetated bank interpreted by these previous workers as a moraine deposited by a cirque glacier is actually an erosional feature, cut by White Brook as it incised into valley fill deposits. The lithology of pebbles (Craft, 1976; Barclay, 1993) and light mineral fractions (Craft, 1976) of tills in the valley suggest deposition from continental ice flowing southwest, up White Brook valley. Furthermore, the interpretation of a till down-valley from

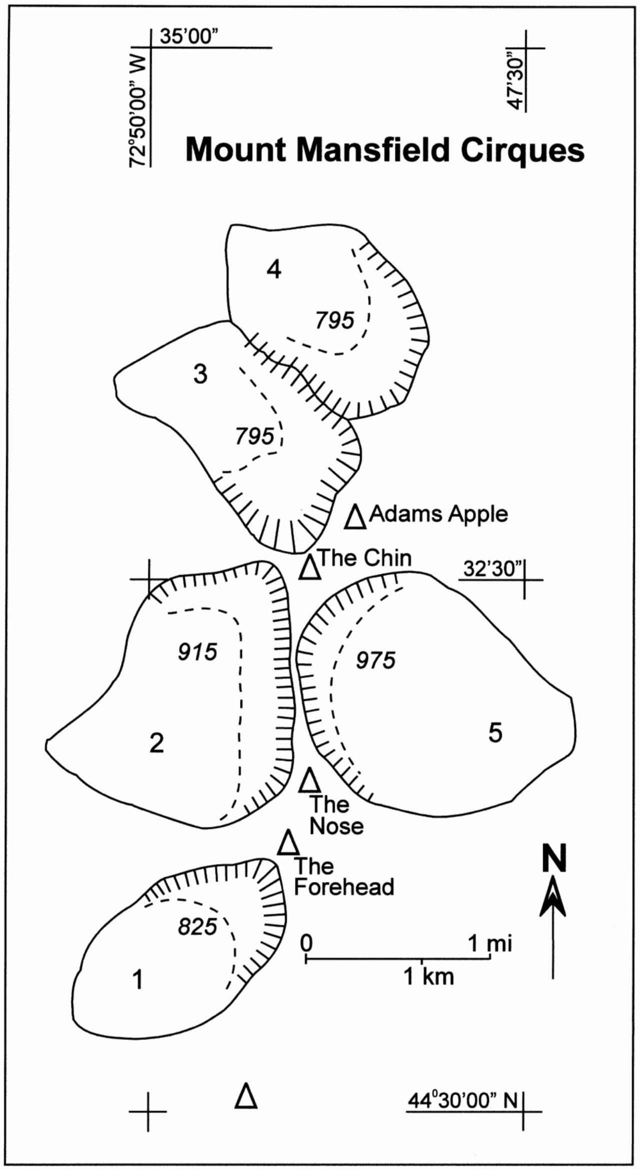


FIGURE 9. Cirque morphometric map for Mount Mansfield, Vermont. Solid lines indicate outlines of cirques, hachured solid lines are cirque headwalls, dashed lines are cirque schrunds, with altitudes indicated in meters. Cirque numbers refer to Table I.

Morphométrie du cirque de vallée du mont Mansfield, au Vermont. Les lignes pleines font le pourtour des cirques, les lignes hachurées montrent les murs de rimaye, les tirets représentent les rimayes (altitude en m). La numérotation des cirques est celle du tableau I.

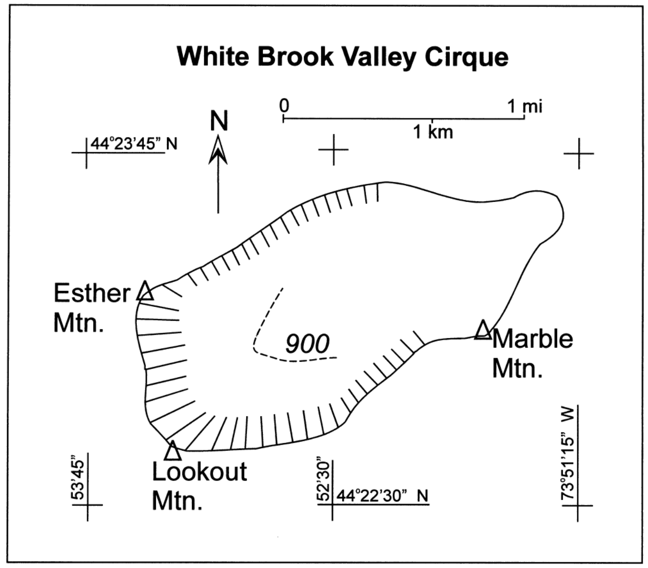


FIGURE 10. Cirque morphometric map for White Brook valley, Adirondack Mountains, New York. Solid line indicates outline of cirque, hachured solid line is cirque headwall, dashed line is cirque schrund, with altitude indicated in meters.

Morphométrie du cirque de la vallée du White Brook, dans les Adirondack (New York). Les lignes pleines font le pourtour des cirques, les lignes hachurées montrent les murs de rimaye et les tirets représentent les rimayes (altitude en m).

the mouth of White Brook valley as a local glacier deposit by Craft (1976) contradicts reconstructions by Franzi (1992) of glacial lake levels in the adjacent Ausable River valley.

Although sediment cores have not been recovered from high-elevation basins in the High Peaks region for radiocarbon dating deglaciation, a sediment core was obtained from Readway Pond at an elevation of 424 m on an outwash plain on the northwestern flank of the Adirondacks. Although this site is far away from any Adirondack cirques, bulk detrital organic material from the 754-764 cm interval of a sediment core yielded a conventional radiocarbon age of 12,640 ± 430 yrs BP (GX-14486), thus providing a minimum-limiting age for continental ice retreat from the Star Lake moraine (Pair and Rodrigues, 1993; Davis *et al.*, 1995a). If the deglaciation model described for the White Mountains in New Hampshire, as well as the Longfellow Mountains and Mount Katahdin in Maine, is also appropriate for deglaciation of the Adirondacks, then cirques in the High Peaks region may have been free of ice as early of 13,000 years ago.

CATSKILL MOUNTAINS, NEW YORK

The Catskill Mountains (Fig. 1) may hold the best evidence for local mountain glaciers following wastage of continental ice. Rich (1906, 1935) identified numerous basins in the Catskills with steep headwalls and side walls, broad floors, and looped end moraines composed of locally-derived till. At the head of Johnson Hollow Brook valley (Schoendorf cirque; Fig. 11), Rich (1906) also identified striae that he interpreted to be formed by a local glacier flowing northeast. Johnson (1917) agreed with all of Rich's (1906) observations except for the striae in Johnson Hollow Brook valley, which he believed were formed by regional flow of continental ice towards the southwest, up the valley. Rich (1935) believed that post-glacial talus deposits caused the cirque headwalls and side walls to be less steep than they were in the past. Cadwell (1986), who concurred with earlier suggestions that a local glacier occupied Schoendorf cirque, suggested that local ice may have persisted long enough to develop a set of glacial terraces on the cirque headwall. Cadwell (1986) also noted a pond

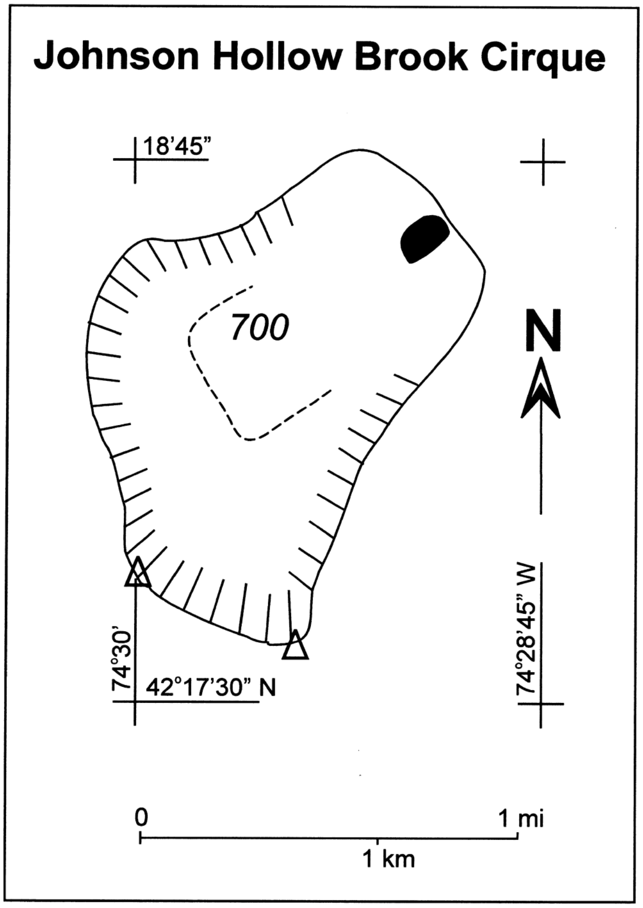


FIGURE 11. Cirque morphometric map for Johnson Hollow Brook, Catskill Mountains, New York. Solid line indicates outline of cirque, hachured solid line is cirque headwall, dashed line is cirque schrund, with altitude indicated in meters. Water body shown in black.

Morphométrie des cirques du Johnson Hollow Brook, dans les Catskill (New York). Les lignes pleines font le pourtour des cirques, les lignes hachurées montrent les murs de rimaye et les tirets représentent les rimayes (altitude en m). L'étendue d'eau est en noir.

dammed by the moraine across the mouth of the cirque, and he suggested that regional deglaciation of continental ice in the Catskills occurred about 15,000 yrs BP.

The moraine-dammed pond in Schoendorf cirque (Fig. 11) attracted the attention of Donald Rodbell and his students at Union College in Schenectady in the middle 1990s. Lederer (1998) and Lederer and Rodbell (1998) confirmed the concave up-valley form of the moraine, determined the moraine's composition to be 98% locally derived clasts, obtained a sediment core from the pond, and conventionally radiocarbon-dated bulk detrital organic material at the 545 cm depth. This depth marked an abrupt transition from pink clay below to organic-rich gyttja above, which was radiocarbon-dated to 10,860 ± 115 yrs BP (GX-23836). If the pink clay is a late-glacial diamict of local origin, as suggested by Lederer (1998), Johnson Hollow Brook valley provides the only radiocarbon age from a cirque in the northeastern United States that suggests persistence of local alpine ice following recession of the last continental icesheet.

THEORETICAL GLACIER PROFILE RECONSTRUCTIONS

Shreve (1985a, b) constructed a theoretical continental ice-surface profile based on esker data to suggest that Mount Katahdin was a nunatak, with ice at the late Wisconsinan maximum reaching only to about the 1100-m altitude on the mountainside. However, field data, such as unweathered erratics in till near the summit areas, and theoretical ice-surface profiles based on Nye's (1952) shear-stress equation

$$\tau_b = \rho g h \sin\alpha \qquad (1)$$

where τ_b is basal shear stress of the ice, ρ is ice density, g is gravity, h is ice thickness, and α is the surface gradient of the ice, suggest that Mount Katahdin was covered by continental ice during the Late Wisconsinan (Davis, 1989). Ackerly (1989) constructed ice-surface profiles for 37 proposed mountain glaciers in northeastern United States using the above shear-stress equation, modified by introducing a shape factor (Nye, 1952) based on the shape of the valley cross section and the ratio of glacier width to depth, such that

$$\tau_b = \rho g h F \sin\alpha \quad \text{and} \quad F > 1 \qquad (2)$$

Given reasonable assumptions for assigned values τ_b and $\sin\alpha$, Ackerly's (1989) reconstructions supported the existence of local alpine glaciers at some time in the past for the Presidential Range, Mount Moosilauke, Mount Katahdin, the Longfellow Mountains, the Catskills, and some sites in the Adirondacks, whether or not these sites were occupied by cirque glaciers subsequent to wastage of continental ice. However, for Wagner's (1970, 1971) proposed sites for cirque glaciers in the Green Mountains, Ackerly (1989) found that reconstructed local ice thickness exceeded the depth of their respective valleys or the altitude of their up-valley cols. He noted that these latter sites were either very shallow basins or very gently sloping valleys. Thus, Ackerly's (1989) theoretical ice-surface profiles support the conclusions of Waitt and Davis (1988), which questioned whether Miller Brook, Ritterbush, and Belvidere valleys were true cirques. Ackerly (1989) did not model ice profiles for the high-altitude cirques on Mount Mansfield (Fig. 9).

CIRQUE MORPHOMETRY

BACKGROUND

Lewis (1938) defined four important characteristics of cirques: 1) steep and usually shattered headwalls and side walls, 2) a gentle rock floor usually with evidence of overdeepening and smoothing, 3) a rock lip or threshold at the mouth, and 4) a rock node at the junction of the headwall and cirque floor. In areas of the world where cirques have undergone substantial post-glacial mass wasting and were last exposed to continental icesheets rather than local glaciers, the latter three characteristics are commonly buried, and therefore difficult to observe, both in the field and even on topographic maps. Evans and Cox (1974, 1995) provide a more quantitative basis for describing cirque morphometry on topographic maps, some of which is adopted in this report.

Although cirques are not as numerous in the northeastern United States as in other mountainous areas of the world (*e.g.*, Andrews, 1965; Andrews and Dugdale, 1971; Williams, 1975; Graf, 1976; Evans, 1977; Evans and Cox, 1995), their altitude, orientation, and form can tell us much about paleoclimatic conditions. Flint (1971, p. 67-70, 133-138) suggested that cirque floor altitudes approximate the orographic snow-line at the time of cirque glacier erosion. In a comparison of six methods for estimating equilibrium-line altitudes (ELAs), Meierding (1982) found that measuring cirque floor altitudes was one of the more rapid methods, albeit one of the more subjective, and was prone to underestimation of ELAs.

Goldthwait (1970) estimated the schrund altitude for former mountain glaciers in the Presidential Range as the elevation at which projections of the steepest headwall long slope and the average cirque floor slope intersect (Fig. 2). In all but two cases for the Presidential Range (Oakes Gulf and Castle Ravine; Goldthwait, 1970, p. 100, Fig. 9), the schrund altitudes determined by this method and a simple measurement of change in spacing between contour lines on topographic maps differed by less than 20 m. Because schrund altitudes are higher than cirque floors, their measurement is a more conservative method for estimating the former depression of ELA than is measuring cirque floor altitudes. Goldthwait's (1970) calculation of a mean summer temperature lowering between 5 and 10°C necessary to support cirque glaciers in the Presidential Range today might be considered a minimum estimate. However, other factors may also be important for estimating paleo-ELAs, such as the amount of winter precipitation or the aspect of cirques, as noted by Goldthwait (1970).

Besides schrund altitude and cirque aspect, other parameters of cirque morphology are also useful for estimating paleoenvironmental conditions of cirque glaciers. For example, cirque length-to-height ratios allow one to estimate glacier surface profiles, which, in turn, could influence erosion potential of cirque glaciers (Embleton and King, 1975, p. 209-210). Cirque length-to-height and length-to-width ratios also allow one to compare degree of cirque development, which can be used to categorize cirques by grade (Evans and Cox, 1995). Cirque volumes can be calculated by: (length × height × width/2) (Andrews, 1975), which allows estimation of duration of cirque glaciation if given empirically-derived cirque erosion rates (Andrews, 1972; Andrews and LeMasurier, 1973; Reheis, 1975; Anderson, 1978). Headwall-to-floor slope ratios allow one to estimate the amount of over-deepening of cirques, although other morphometric methods are probably more accurate (Haynes, 1968).

METHODS

Morphometric measurements for cirques in seven different areas of the northeastern United States as measured from topographic maps, most published since 1988, are summarized in Table I. Specific names, scales, and contour intervals for these maps are provided in the Appendix 1. Cirque grade, a qualitative evaluation of cirque form, follows the classification of Evans and Cox (1995). Aspect is measured as the direction faced by the central headwall, with the headwall orientation defined as a perpendicular to the long axis of each cirque; negative values increase from 360° to 180° to facilitate averaging. Schrund altitudes are measured as the most obvious break in slope between cirque headwalls and floors as seen by a change in spacing between contour lines (Fig. 2); these provide slightly higher estimates than the method of Goldthwait (1970). Numerous measurements of slope between tops of headwalls and schrund altitudes provide average headwall slopes somewhat lower than values for steepest long slopes of headwalls as measured by Goldthwait (1970). Average floor inclinations below schrund altitudes are measured in similar fashion to Goldthwait (1970).

PRESIDENTIAL RANGE, NEW HAMPSHIRE

Four basins in the Presidential Range in addition to those summarized by Goldthwait (1970) have been identified as cirques in this report: Ammonoosuc, Burt, and Cascades Ravines and "Franklin Basin" (Table I, Figs. 1, 3). These four cirques, as well as "Sphinx Basin" and "Monroe Basin" recognized by Goldthwait (1970), are all ranked grade 4, or poor in the classification of Evans and Cox (1995). Although there may be some doubt whether these basins should be classified as cirques, well-developed characteristics (namely steep headwalls and side walls) compensate for weak ones (namely lack of broad and/or gently sloping floors). Only four cirques in the Presidential Range ranked grade 1 (Upper Great Gulf, Huntington, Upper Tuckerman, and Lower Tuckerman Ravines); King Ravine and Upper Oakes Gulf were ranked only grade 2 because of their steeply sloping floors and weakly developed side walls, respectively. Goldthwait (1970) suggested that snow drifted by prevailing southwesterly winds during glacial periods was important in determining locations of cirques in the Presidential Range; this suggestion is supported by the northern and eastern aspects of the six cirques graded 1 and 2. The mean aspect for all Presidential Range cirques is about 53° azimuth, with a standard deviation of 88° (Table I). The average schrund altitude of 1270 m for the six cirques graded 1 and 2 is about 40 m higher than the average schrund altitude of 1230 m for all Presidential Range cirques (Table I). This difference could indicate greater duration of cirque glacier erosion in basins with higher altitudes or more recent occupation by cirque glaciers. However, the schrund altitudes labeled "RPG" in Table I (from Goldthwait, 1970, Table 1, p. 89) suggest grade 1 and 2 cirques are only about 25 m higher than the average schrund altitude for all cirques, about 1245 m.

From morphometric data on height, width, and length of cirques (Table I), length-to-height (L:H) and length-to-width (L:W) ratios and cirque volumes are calculated. The mean

TABLE I

Cirque morphometric characteristics, New England area, U.S.A.

Cirque Name	Cirque Grade[1]	Cirque Aspect (°)[2]	Schrund Altitude (m)[3]	Schrund Altitude (m)[3]	Aver. Height (m)[4]	Aver. Width (m)[5]	Cirque Length (m)[6]	Average headwall slope (°)[7]	Aver. floor inclination (°)[8]	Length: Height Ratio	Length: Width Ratio	Headwall: Floor Slope Ratio	Cirque Volume (km^3)[9]
Presidential Range, N.H.				**RPG**									
1. Ammonoosuc Ravine	4	–80	1080		530	1370	1525	30	10	2.88	1.11	3.00	0.55
2. Burt Ravine	4	–95	1170		630	990	1830	27	13	2.90	1.85	2.08	0.57
3. Castle Ravine	3	–45	1095	1067	585	840	1525	30	13	2.61	1.82	2.31	0.37
4. Cascade Ravine	4	–45	990		435	610	1065	29	11	2.45	1.75	2.64	0.14
5. King Ravine	2	–20	1230	1165	660	915	1525	36	17	2.31	1.67	2.12	0.46
6. Bumpus Basin	3	5	990	964	360	535	1065	27	10	2.96	1.99	2.70	0.10
7. Madison Gulf	3	120	1260	1214	420	535	1145	32	13	2.73	2.14	2.46	0.13
8. Jefferson Ravine	3	105	1260	1262	570	990	1675	38	11	2.94	1.69	3.45	0.47
9. "Sphinx Basin"	4	100	1470	1409	270	535	610	31	21	2.26	1.14	1.48	0.04
10. Upper Great Gulf	1	10	1350	1342	650	1065	2135	42	9	3.28	2.00	4.67	0.74
11. Huntington Ravine	1	135	1260	1287	630	610	1525	43	11	2.42	2.50	3.91	0.29
12. Upper Tuckerman Ravine	1	110	1365	1378	285	535	610	38	11	2.14	1.14	3.45	0.05
13. Lower Tuckerman Ravine	1	50	1200	1232	435	685	915	27	6	2.10	1.34	4.50	0.14
14. Gulf of Slides	3	65	1280	1262	340	1065	990	30	12	2.91	0.93	2.50	0.18
15. Upper Oakes Gulf	2	160	1230	1275	385	915	1415	32	3	3.68	1.55	10.67	0.25
16. "Monroe Basin"	4	165	1320	1342	310	685	915	35	9	2.95	1.34	3.89	0.10
17. "Franklin Basin"	4	160	1340		340	455	990	33	17	2.91	2.18	1.94	0.08
average	**2.8**	**52.9**	**1228.8**	**1246.1**	**452.5**	**751.4**	**1222.5**	**33.6**	**11.2**	**2.70**	**1.64**	**3.62**	**0.25**
standard deviation	**1.2**	**87.8**	**131.3**	**124.1**	**137.6**	**257.4**	**424.7**	**5.0**	**4.2**	**0.42**	**0.44**	**2.08**	**0.21**
												total volume	**4.21**
Mt. Moosilauke, N.H.				**GMH**									
1. Jobildunk Ravine	2	140	1135	1130	260	950	1950	23	5	7.50	2.05	4.60	0.24
2. Gorge Brook Ravine	3	150	1025	1220	305	500	1440	22	7	4.72	2.88	3.14	0.11
3. Benton Ravine	3	–30	1035	1100	610	750	1780	32	12	2.92	2.37	2.67	0.41
4. Little Tunnel Brook Ravine	3	–10	730		425	600	1550	23	9	3.65	2.58	2.56	0.20
average	**2.8**	**62.5**	**981.3**	**1150.0**	**400.0**	**700.0**	**1680.0**	**25.0**	**8.3**	**4.70**	**2.47**	**3.24**	**0.24**
standard deviation	**0.5**	**95.7**	**174.7**	**62.4**	**156.4**	**195.8**	**229.1**	**4.7**	**3.0**	**2.01**	**0.35**	**0.94**	**0.12**
												total volume	**0.96**
Mount Katahdin, Me.				**PTD**									
1. Klondike Basin	1	–90	1065	1052	215	455	1145	26	2	5.33	2.52	13.00	0.06
2. Upper Northwest Basin	1	–45	915	899	275	535	1600	38	8	5.82	2.99	4.75	0.12
3. Lower Northwest Basin	2	–45	915		365	760	1370	39	9	3.75	1.80	4.33	0.19
4. Little North Basin	2	90	795	808	305	535	1220	25	5	4.00	2.28	5.00	0.10
5. North Basin	1	90	1135	1067	510	990	1980	39	8	3.88	2.00	4.88	0.50
6. Great Basin	2	85	1065	1052	440	1450	1525	39	8	3.47	1.05	4.88	0.49
7. South Basin	1	15	945	945	650	1065	1675	47	4	2.58	1.57	11.75	0.58
8. Witherle Ravine	3	–50	1065	1143	475	535	1450	21	15	3.05	2.71	1.40	0.18
average	**1.6**	**6.3**	**987.5**	**995.1**	**404.4**	**790.6**	**1495.6**	**34.3**	**7.4**	**3.98**	**2.12**	**6.25**	**0.28**
standard deviation	**0.7**	**73.7**	**112.9**	**115.7**	**142.3**	**350.4**	**266.0**	**9.1**	**3.9**	**1.09**	**0.64**	**3.97**	**0.21**
												total volume	**2.21**

TABLE I (continued)

Cirque morphometric characteristics, New England area, U.S.A.

Longfellow Mountains, Me.				**HWB/PEC**									
1. Sugarloaf Mountain	4	−65	795	792	320	535	1120	21	11	3.50	2.09	1.91	0.10
2. Sugarloaf Mountain	4	170	760	762	440	915	1980	25	5	4.50	2.16	5.00	0.40
3. Sugarloaf Mountain	4	165	825	713	470	915	1600	30	7	3.40	1.75	4.29	0.34
4. Sugarloaf Mountain	4	75	795	689	440	1065	1675	31	6	3.81	1.57	5.17	0.39
5. Crocker Mountain	4	75	885	853	425	760	1450	27	7	3.41	1.91	3.86	0.23
6. Crocker Mountain	3	80	915	847	275	760	1220	22	7	4.44	1.61	3.14	0.13
7. Crocker Mountain	4	−85	900	817	425	1370	2590	27	6	6.09	1.89	4.50	0.75
8. Crocker Mountain	4	−10	825	725	395	760	1370	22	8	3.47	1.80	2.75	0.21
Black Nubble	4	−125	865	860	290	990	2135	19	4	7.36	2.16	4.75	0.31
Tim Mountain	4	−120	795	750	350	1850	2745	18	3	7.84	1.48	6.00	0.89
average	**3.9**	**16.0**	**836.0**	**780.8**	**383.0**	**992.0**	**1788.5**	**24.2**	**6.4**	**4.78**	**1.84**	**4.14**	**0.37**
standard deviation	**0.3**	**111.9**	**52.3**	**62.1**	**69.2**	**375.1**	**560.0**	**4.5**	**2.2**	**1.70**	**0.24**	**1.24**	**0.26**
												total volume	**3.75**
Green Moutains, Vt.				**RBW/PTD**									
Miller Brook valley	3	135	395	360	715	1675	3505	25	2	4.90	2.09	12.50	2.10
Ritterbush valley	3	115	360	335	220	1830	2135	18	2	9.70	1.17	9.00	0.43
Belvidere valley	3	−105	360	365	190	1525	2440	11	1	12.84	1.60	11.00	0.35
1. Mansfield	3	−125	825	1005	610	950	2200	25	12	3.61	2.32	2.08	0.64
2. Mansfield	4	−105	915	945	640	1900	2400	29	11	3.75	1.26	2.64	1.46
3. Mansfield	3	−50	795	795	760	1100	2300	27	13	3.03	2.09	2.08	0.96
4. Mansfield	3	−55	795	795	605	1200	2150	24	12	3.55	1.79	2.00	0.78
5. Mansfield	3	110	975	975	700	1700	2450	37	13	3.50	1.44	2.85	1.46
Mansfield 1 - 5 average	**3.2**	**−45.0**	**861.0**	**903.0**	**663.0**	**1370.0**	**2300.0**	**28.4**	**12.2**	**3.49**	**1.78**	**2.33**	**1.06**
standard deviation	**0.4**	**92.4**	**80.5**	**100.8**	**66.1**	**408.7**	**127.5**	**5.2**	**0.8**	**0.27**	**0.44**	**0.39**	**0.38**
												total volume	**5.30**
Adirondack Mountains, N.Y.													
White Brook valley	3	65	900		325	1250	3050	22	9	9.38	2.44	2.44	0.62
Catskill Mountains, N.Y.													
Johnson Hollow Brook valley	2	50	700		370	1220	1980	24	7	5.35	1.62	3.43	0.45

[1] Grade follows classfication of Evans and Cox (1995), whereby 1 = classic, with all textbook attributes, 2 = well-defined, with headwall and floor clearly developed and headwall curves around cirque floor, 3 = definite, with no debate over cirque status, but one characteristic may be weak, 4 = poor, some doubt, but well-developed characteristics compensate for weak ones, 5 = marginal, with cirque status and origin doubtful.
[2] Aspect is direction faced by central headwall perpendicular to long axis of cirque measured to nearest 5° azimuth (negative values increase from 360° to 180°).
[3] Schrund altitudes in left column measured as most obvious break in slope denoted by contour lines, following method of Evans and Cox (1995), whereby headwall slopes generally more than 35° and floor slopes generally less than 20°, with limit between the two at 27°; schrund altitudes in right column measured by method of Goldthwait (1970), with initials of workers for particular areas as explained in text.
[4] Height measured from average top of headwall to lowest floor altitude (to nearest 5 m).
[5] Average width determined by numerous measurements from top of sidewall to top of opposite sidewall along long axis of cirque (to nearest 5 m).
[6] Length measured from top of headwall to cirque mouth, or where sidewalls abruptly end or drop in altitude (to nearest 5 m).
[7] Average headwall slope measured from top of headwall to cirque mouth, or where sidewalls abruptly end or drop in altitude (to nearest 5 m).
[8] Average floor inclination measured below schrund altitude.
[9] Volume calculated by: (height × width × length) / 2.

L:H ratio of 2.70:1 for Presidential Range cirques (n = 17) compares with averages of 2.8:1 to 3.2:1 for over 400 cirques in English Lake District (Embleton and King, 1975) and a median of 4.29:1 for 165 cirques on Baffin Island (Andrews and Dugdale, 1971). The mean L:W ratio of 1.64:1 for Presidential Range cirques compares with a median of 1.3:1 for the same 165 cirques on Baffin Island. The average volume of material removed from Presidential Range cirques is 0.25 km^3, with a total 4.25 km^3. Using a range of empirically-derived cirque erosion rates (Anderson, 1978), about 5.9×10^4 to 10.4×10^6 yrs would be required to erode the average Presidential Range cirque, assuming that 100% of the cirque forms are created by glacial erosion.

Average headwall slopes summarized in this report (mean 33.6°, Table I) are naturally lower than the *steepest* headwall slope segments (mean 37°) measured for the Presidential Range by Goldthwait (1970, Table 1, p. 89), although the average inclinations of floors below schrund altitudes are similar (11.2° vs 10°, respectively). Only one cirque floor (“Sphinx Basin”) has an inclination (21°) higher than the upper limit (20°) suggested by Evans and Cox (1995); however, the majority of Presidential Range cirques have average headwall slopes a few degrees less than their suggested lower limit (35°). All Presidential Range cirques have headwall-to-floor slope ratios below 4.7:1, except for Upper Oakes Gulf (10.67:1; average floor inclination 3°); the standard deviation of 2.08 drops in half if the Upper Oaks Gulf ratio is not included. Thus, using new maps and different methods for measuring schrund altitudes and headwall slopes do not result in headwall-to-floor slope ratios much different than those determined by Goldthwait (1970) for Presidential Range cirques.

MOUNT MOOSILAUKE, NEW HAMPSHIRE

Morphometric data for four cirques on Mount Moosilauke (Figs. 1, 4) in the western White Mountains of New Hampshire are summarized in Table I. In comparison with cirques in the Presidential Range, cirques on Mount Moosilauke average the same grade (2.8), have a similar high variability in aspect (two with azimuths southeast; two with azimuths north-northeast), average almost 250 m lower in schrund altitudes (although average about 165 m lower if only highest three cirques on Mount Moosilauke considered), have higher length-to-height and length-to-width ratios (4.40:1 vs. 2.70:1 and 2.47:1 vs. 1.64:1, respectively), have lower average headwall and floor slope angles but similar average headwall-to-floor slope ratios (3.24 vs. 3.62), and have almost identical average volumes (0.24 vs. 0.25 km^3 per cirque). Measured schrund altitudes for the three highest cirques on Mount Moosilauke are about 85 m lower (1065 m vs. 1150 m) than measured by Haselton (1975).

MOUNT KATAHDIN, MAINE

Morphometric data for eight cirques on Mount Katahdin (Figs. 1, 5) in west-central Maine are summarized in Table I. In comparison with cirques in the Presidential Range, those on Mount Katahdin average higher in grade (1.6 vs. 2.8, with four ranked grade 1), have a similar high variability in aspect (four with azimuths west or northwest; four with azimuths east or northeast), average over 240 m lower in schrund altitudes, have higher average length-to-height and length-to-width ratios (3.98:1 vs. 2.70:1 and 2.12:1 vs. 1.64:1, respectively), have very similar average headwall slopes (34.3° vs. 33.6°) but a higher average headwall-to-floor slope ratio (6.25:1 vs. 3.62:1), and have similar average volumes (0.28 vs. 0.25 km^3 per cirque, albeit a 0.21 km^3 standard deviation for each group), with a total cirque volume about half of the Presidential total, reflecting about half the number of cirques.

The slightly different schrund altitudes for Katahdin’s cirques in the right column labeled “PTD” in Table I were measured from a 15-minute topographic map (Davis, 1976), as opposed to measurements in the left column that were made on a 7.5-minute map; however, the contour interval of 20 ft (6 m) was the same on both maps. One might be tempted to explain the lower schrund altitudes on Mount Katahdin as indication of greater snow-line depression than in the Presidential Range; however, different underlying bedrock (Devonian granite on Mount Katahdin [Osberg *et al.*, 1985] vs. Silurian-Devonian schist, gneiss, and quartzite of the Rangeley, Perry Mountain, Small Falls, Madrid, and Littleton Formations in the Presidential Range [Eusden *et al.*, 1996]) might also be an important controlling factor for schrund altitude (*e.g.*, Evans, 1994). In Mount Katahdin’s three large east-facing cirques, the steep headwalls appear to be controlled by the prominent vertical jointing in the granite bedrock and the consequent release of large exfoliation sheets.

LONGFELLOW MOUNTAINS, MAINE

Morphometric data for 10 cirques in the Longfellow Mountains in west-central Maine (Figs. 1, 6) are summarized in Table I. In comparison with cirques of the Presidential Range, Mount Moosilauke, and Mount Katahdin, cirques in the Longfellow Mountains average lower in grade, with nine of 10 cirques grading 4. Cirque aspect in the Longfellow Mountains is more variable and schrund altitudes average about 390, 150, and 160 m lower than in the Presidential Range, on Mount Moosilauke, and on Mount Katahdin, respectively. Schrund altitudes measured from 7-5 minute topographic maps average about 55 m higher than schrund altitudes measured by Borns and Calkin (1977) from 15-minute maps (labeled “HWB/PEC” in Table I), although the contour intervals were the same at 20 ft (6 m).

Although average height is smaller, average width and length are greater, hence average cirque volume is greater for Longfellow than for Presidential, Moosilauke, or Katahdin cirques. Average headwall slopes are less steep than those for cirques of the Presidential Range and Mount Katahdin, but similar to those on Mount Moosilauke. Concomitant lower floor inclinations result in an average headwall-to-floor slope ratio between that for cirques of the Presidential Range, Mount Moosilauke, and Mount Katahdin (4.14:1 vs. 3.62:1, 3.24:1, and 6.25:1, respectively). There does not appear to be a relationship between cirque morphometry and bedrock in the Longfellow Mountains (Table I), as cirques with similarly steep headwalls are cut into a variety of bedrock types (Osberg *et al.*, 1985), including Devonian granodiorite (Crocker 6 and 7 and Black Nubble cirques), Devonian gabbro and ultramafics (Crocker 5 and 8 and Sugarloaf 1 cirques), and pelites of the Silurian Perry Mountain and Rangeley Formations (Sugarloaf 2, 3, and 4, and Tim Mountain cirques).

GREEN MOUNTAINS, VERMONT

Waitt and Davis (1988, Table 2. p. 508) summarized data for two groups of cirques in the Green Mountains (Fig. 1), one group consisting of three low-altitude basins (Figs. 7, 8), the other seven high-altitude cirques in the Mount Mansfield area (Fig. 9). Of the seven high-altitude cirques, Waitt and Davis (1988) considered two to be questionable in origin, so data are only presented here for the five cirques on Mount Mansfield (Fig. 9, Table I). Because the three low-altitude basins led Wagner (1970, 1971) and Connally (1971) to suggest that local glaciers persisted in the Green Mountains following dissipation of continental ice, these are considered first. These three basins are unusual in that two hold very large lakes compared with the typical size of tarns, two are oriented to the southwest, they average about 370 m in schrund altitudes, one has a volume almost three times larger than any other cirque in the northeastern United States, and headwalls are about half the slope angle of cirques in the Presidential Range and Mount Katahdin. These characteristics, along with odd digitate shapes (Figs. 7, 8), led Waitt and Davis (1988) to question whether the basins were cirques at all, but rather to interpret them as valley heads last occupied by late-glacial tongues of continental ice.

Morphometric data for the five high-altitude cirques on Mount Mansfield in the northern Green Mountains (Figs. 1, 9) are summarized in Table I. In comparison with cirques of the Presidential Range, Mount Moosilauke, Mount Katahdin, and the Longfellow Mountains, the high-altitude cirques on Mount Mansfield are similar in grade, with an average of 3.2. Four of the high-altitude Mansfield cirques have a westerly aspect, with an average of northwest for the five cirques. Schrund altitudes average about 55 m higher than in the Longfellow Mountains, but about 270, 145, and 150 m lower than cirques in the Presidential Range and Mount Moosilauke, and Mount Katahdin, respectively. With the exception of Mansfield cirque #1, earlier measurements of schrund altitudes reported in Waitt and Davis (1988; labeled “RBW/PTD” in Table I) are similar to the new measurements in Table I. Most other morphometric measurements for the high-altitude cirques on Mount Mansfield are similar to those for cirques of the Presidential Range, Mount Moosilauke, Mount Katahdin, and the Longfellow Mountains; however, the Mansfield cirques are 3 to 4 times larger in average volume. In summary, the high-altitude cirques on Mount Mansfield appear to have morphometric characteristics that are similar to those of other cirques in northern New England, but have less in common with the three low-altitude cirques or valley heads in the northern Green Mountains.

ADIRONDACK MOUNTAINS, NEW YORK

Craft (1976) identified 224 cirque basins in the Adirondacks and proposed a number of sites as locations for cirque glaciers persisting after wastage of continental ice. Warburton (1982) noted that many cirques in the Adirondacks have breached headwalls, whereas other basins have been so severely eroded by valley glaciers that they have lost much of their cirque form, a likely indication that erosive continental ice post-dated cirque glaciation in the High Peaks region. Based on field mapping, Barclay (1993) rejected a number of Craft’s (1976, 1979) sites as potential locations of cirque glaciers post-dating continental ice recession; however, Donald Rodbell (oral commun., 1998) suggests that many cirques in the Adirondacks have moraines, including some with tarns that could yield sediment cores and minimum-limiting radiocarbon ages for deglaciation.

Morphometric data for only one Adirondack site, White Brook valley on the northeast side of Whiteface Mountain (Figs. 1, 10), is presented in Table I. This grade 3 cirque has a northeast aspect, a schrund altitude between the average schrund altitudes for cirques in the Longfellow Mountains and on Mount Katahdin, and a greater length than the average cirque in the Presidential Range, on Mount Katahdin, and in the Longfellow Mountains. However, the average headwall slope is only 22°, similar to the low average headwall slope angles in the Longfellow and Green Mountains.

CATSKILL MOUNTAINS, NEW YORK

Johnson Hollow Brook valley is one of many cirques in the Catskills proposed to be sites for local glaciers that flowed independently after recession of continental ice (Rich, 1906, 1935; Johnson, 1917). Because Johnson Hollow Brook valley (Figs. 1, 11) is oriented toward the northeast, it is ideally suited to maximize wind drifted snow, to minimize solar radiation, and to provide ice-flow indicators opposite those of continental ice. The cirque is similar in size to the larger cirques in the White Mountains, on Mount Katahdin, and in the Longfellow Mountains; however, its average headwall slope (24°) is about 10° lower and its schrund altitude (700 m) is between 300 and 500 m lower than the average cirque on Mount Katahdin and in the Presidential Range, respectively.

DISCUSSION

Using an atmospheric lapse rate (6°C/1km; *e.g.*, Barry, 1992, p 45) and assuming a current July freezing isotherm at 3050 m altitude in northern New England, Goldthwait (1970) and Davis (1976) estimated that cirque glaciers in the Presidential Range and on Mount Katahdin would have required summer temperature depressions about 9°C and 13°C below modern-day values, respectively. Using a 5.3°C/1km lapse rate, Loso *et al.* (1998) calculated a summer temperature depression about 13.6°C below modern-day values to support a local glacier in the Miller Brook valley of the Green Mountains, assuming an equilibrium-line altitude of 480 m (using the 6°C/1km lapse rate and the 395 m schrund altitude measured in Table I would yield a summer temperature depression about 16°C). Using an atmospheric lapse rate of 6°C/1km and assuming a current July freezing isotherm at 3350 m altitude in southern New York would require a summer temperature depression about 16°C below modern-day values to support a cirque glacier at the head of Johnson Hollow Brook valley in the Catskills. However, these estimates do not consider increased winter precipitation (Leonard, 1989), which could compensate for part of the required temperature depression. For example, Lederer (1998) suggested that winter precipitation between 12,000 and 11,000 yrs BP may have increased in the Catskills because of the proximity of glacial Lake Iroquois (Pair and Rodrigues, 1993). Moreover, favorable cirque aspect for shielding from summer insolation and enhancement of wind-drifted snow from prevailing winds (Goldthwait, 1970) also may have influenced snow accumulation in cirques. Indeed, Havens (1960) noted that large snow banks commonly remained throughout the summer on the floor of Upper Tuckerman Ravine in the Presidential Range during the early 1900s. However, lichenometric data, similar to that collected on the summit of Mount Washington by Mayewski and Jeschke (1978), suggest that snow banks in Tuckerman Ravine did not persist or increase in size sufficiently to form a cirque glacier during Neoglaciation (the past 2000 to 3000 years).

If cirque glaciers did persist after recession of continental ice in the northeastern United States, the most likely time for their occurrence would have been during the Younger Dryas (YD) cooling event (Alley *et al.*, 1993; Mayewski *et al.*, 1993), which is radiocarbon-dated about 11,000 to 10,000 yrs BP. The earlier Killarney Oscillation, a cooling event radiocarbon-dated between 11,300 and 11,000 yrs BP (Levesque *et al.*, 1994), was probably too short-lived to support cirque glaciers and much of the northeastern United States remained under continental ice during the Heinrich I cooling event (Bond *et al.*, 1992) about 14,000 yrs BP. Moraines that may have been formed by cirque glaciers are not present in the Presidential Range, Longfellow Mountains, or Adirondacks. Moraines that flank the east and south slopes of Mount Katahdin were formed by continental ice surrounding the mountain, and not by cirque glaciers (Davis, 1976, 1978, 1983, 1989). Radiocarbon ages from pond and bog-bottoms on Mount Katahdin are far too young to provide closely-limiting ages for these moraines (Davis and Davis, 1980); however, one preliminary set of cosmogenic ^{10}Be and ^{26}Al exposure ages for a boulder on one of the moraines suggests an age older than the YD (Davis and Bierman, unpublished data). Cosmogenic radionuclides have provided late-glacial moraine chronologies in many areas of the world (*e.g.*, Gosse *et al.*, 1995a, b). Although most of these areas occur at higher altitudes where cosmogenic radionuclide production rates are higher than in the northeastern United States, the sensitivity of the method is improving and can already resolve late-glacial chronologies at sea level (*e.g.*, Davis *et al.*, 1999).

A moraine-like ridge in the Miller Brook valley in the Green Mountains (Wagner, 1970, 1971; Connally, 1971) is now believed to be the result of late-glacial tongues of continental ice (Waitt and Davis, 1988; Loso *et al.*, 1998) or to be an esker (Wright et al., 1997a, b). Moreover, the cross-valley moraines in Ritterbush valley thought to be as young as 10,800 yrs BP (Sperling *et al.*, 1989) are now known to be more than 12,000 year old, or pre-YD in age, based on radiocarbon ages of sediments from Ritterbush Pond (Bierman *et al.*, 1997; Lin *et al.*, 1995; Lini *et al.*, 1995).

At Johnson Hollow Brook valley in the Catskills, a moraine at the mouth of Schoendorf cirque, long held to be deposited by a local glacier persisting after dissipation of continental ice (Rich, 1906, 1935; Johnson, 1917), was recently shown to be as young as 10,860 yrs BP (Lederer, 1998; Lederer and Rodbell, 1998). Thus, the Catskills stand alone as a location in the northeastern United States where cirque glaciers post-date recession of continental ice, and possibly formed during the early part of the YD. The nearest areas where glacial geological data provide similar evidence for local glaciation post-dating wastage of continental ice are the Maritime Provinces of Canada, where lowland ice expanded during the YD (Stea and Mott, 1989). In northern Maine, lithological analyses and radiocarbon-dating of lake sediment cores also suggest that a local lowland icecap may have persisted into YD time (Dorion, 1998). If local ice were present in Catskill cirques following dissipation of continental ice, one must ask why we do not see evidence for local glaciers surviving continental ice recession in higher-altitude and higher latitude cirques in Maine and New Hampshire.

In the Maritime Provinces, the climatic signal for the YD is pronounced in a variety of lake sediment proxy records, including loss on ignition, sediment grain size, pollen, plant macrofossils, and chironomids (Mayle *et al.*, 1993a, b; Levesque *et al.*, 1994, 1997). In the northeastern United States, the climatic signal for the YD is less pronounced but is still recognizable in LOI, pollen, and macrofossil records from lake sediment cores in northwestern Maine and northern New Hampshire (Thompson *et al.*, 1996, this volume; Dorion, 1998). Elsewhere in the northeastern United States, the YD signal is more subtle in lake sediments, but is nevertheless present in most pollen records (Peteet *et al.*, 1990). For example, in pollen records from lake sediment cores on the western flank of the Adirondacks (Davis *et al.*, 1995a), in southern Vermont (Davis *et al.*, 1995b), in northern Vermont (Lin *et al.*, 1995; P.T. Davis *et al.*, unpublished data), and in northwestern New Hampshire (Likens and Davis, 1975; Davis *et al.*, 1980; Davis and Ford, 1982), a decrease is seen in oak pollen with a concomitant increase in alder pollen during an interval that is radiocarbon dated between about 11,000 and 10,000 yrs BP. To interpret such paleoclimatic signals, one must sample at a resolution high enough to recognize short-lived events, which is now underway for plant macrofossils, pollen, chironomids, and other proxies in many high-altitude ponds in the White Mountains (Ray Spear and Les Cwynar, oral commun., 1998). Unfortunately, few cirques in the northeastern United States are blessed with tarns suitable for such paleoclimatic studies.

CONCLUSIONS

Morphometric analysis suggests that cirques in the Presidential Range in New Hampshire and Mount Katahdin in Maine should be the most likely settings for survival or rejuvenation of local glaciers after icesheet recession. However, evidence for cirque glaciers persisting after wastage of continental ice remains lacking for the Presidential Range and Mount Moosilauke of New Hampshire, Mount Katahdin and the Longfellow Mountains of Maine, and the Adirondack Mountains of New York. Radiocarbon dating of lake sediment cores in the northern Green Mountains of Vermont and in the Catskill Mountains of New York has produced additional chronological data that address the presence of cirque glaciers following dissipation of continental ice in the northeastern United States.

At Ritterbush Pond in northern Vermont, radiocarbon ages from sediment cores suggest that a low-altitude valley head located up-valley of a series of cross-valley moraines was ice-free by 11,940 ^{14}C yrs BP, thus precluding a glacier advance during the Younger Dryas cold event. In Miller Brook valley in northern Vermont, a cross-valley ridge that was thought to be a moraine formed by a cirque glacier is now believed to be an esker. Although some workers argue that these ridges in Vermont are evidence for cirque glaciation, morphometric analysis that includes an average schrund altitude of about 370 meters for these low-elevation valley heads suggests that these landforms were more likely formed by a tongue of continental ice during Late Wisconsinan deglaciation.

At Johnson Hollow Brook valley in the Catskills of New York, a radiocarbon age from basal sediments in a tarn dammed by a moraine suggests that the cirque was not ice-free until 10,860 ^{14}C yrs BP, thus allowing the possibility of a local glacier persisting during the earliest part of the Younger Dryas cold interval. However, cirque morphometric data suggest that a schrund altitude of 700 meters for the cirque at the head of Johnson Hollow Brook valley is much less conducive for persistence of a local glacier than are cirques with similar aspect and much higher schrund altitudes in the Presidential Range and on Mount Moosilauke in New Hampshire and on Mount Katahdin in Maine. This dichotomy can only be resolved by further coring and radiocarbon dating of sediments in ponds and bogs in alpine areas of the northeastern United States.

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APPENDIX 1

TOPOGRAPHIC MAPS USED FOR CIRQUE MORPHOMETRIC ANALYSES

Presidential Range, New Hampshire

All cirque morphometric measurements were made from Bradford Washburn's *Mount Washington and the heart of the Presidential Range, New Hampshire*, 1:20,000-scale topographic map, with a contour interval of 50 feet and intermediate contours of 25 feet, surveyed and edited between 1978 and 1987, produced by Boston's Museum of Science in 1988, published and distributed by the Appalachian Mountain Club, Boston, Mass. 02108. With thousands of laser theodolite survey measurements, this map is far more accurate than any other topographic maps for the Presidential Range. Contour lines are easier to read on later editions, in comparison with the first edition, which used a dark green background pattern. Morphometric measurements of Presidential Range cirques #1-16 (Table I) were also made from the U.S. Geological Survey's *Mt Washington, New Hampshire* 7.5- × 15-minute, 1:25,000-scale metric topographic map, with a contour interval of 6 meters. This map was compiled by photogrammetric methods from aerial photographs taken in 1972, field checked in 1975, and edited in 1982. Morphometric measurements of Presidential Range cirques #14-17 were also made from the U.S. Geological Survey's *Stairs Mtn. Quadrangle, New Hampshire*, provisional edition 7.5-minute, 1:24,000-scale topographic map, with a contour interval of 40 feet. This map was compiled from aerial photographs taken in 1981, field checked in 1983, and edited in 1987.

Mount Moosilauke, New Hampshire

All cirque morphometric measurements were made from the U.S. Geological Survey's *Mt. Moosilauke Quadrangle, New Hampshire*, 7.5-minute, 1:24,000-scale topographic map, with a contour interval of 40 feet. This map was compiled from aerial photographs taken in 1964 and field checked and edited in 1967.

Mount Katahdin, Maine

All cirque morphometric measurements were made from the U.S. Geological Survey's *Mount Katahdin Quadrangle, Maine*, provisional edition 7.5-minute, 1:24,000-scale topographic map, with a contour interval of 20 feet. This map was compiled from aerial photographs taken in 1982, field checked in 1985, and edited in 1988.

Longfellow Mountains, Maine

Morphometric measurements for cirques #1-6 (Table I) were made from the U.S. Geological Survey's *Sugarloaf Mtn., Maine, Quadrangle*, provisional edition 7.5-minute, 1:24,000-scale topographic map, with a contour interval of 20 feet. Measurements for cirques #5-8 and Black Nubble cirque (Table I) were made from the U.S. Geological Survey's *Black Nubble, Maine, Quadrangle*, provisional edition 7.5-minute, 1:24,000-scale topographic map, with a contour interval of 20 feet. These maps were compiled from aerial photographs taken in 1985, field checked in 1987, and edited in 1989. Measurements for Tim Mountain cirque (Table I) were made from the U.S. Geological Survey's *Tim Mountain, Quill Hill, Black Mountain, and Kennebago Lake, Maine Quadrangles*, 7.5-minute, 1:24,000-scale topographic maps, with a contour interval of 20 feet. These maps were compiled from aerial photographs taken in 1966, and field checked and edited in 1969 and 1970.

Green Mountains, Vermont

Morphometric measurements for Miller Brook valley cirque were made from the U.S. Geological Survey's *Bolton Mountain, Vt., Quadrangle*, 7.5-minute, 1:24,000-scale topographic map, with a contour interval of 20 feet. This map was compiled from aerial photographs taken in 1947, field checked and edited in 1948, and photo-inspected in 1983. Morphometric measurements for Belvidere and Ritterbush valley cirques were made from the U.S. Geological Survey's *Eden and Hazens Notch, Vermont, Quadrangles*, provisional edition 7.5-minute, 1:24,000-scale topographic maps, with a contour interval of 6 meters. These maps were compiled from aerial photographs taken in 1980 and 1981, field checked in 1982, and edited in 1986. Morphometric measurements for the Mount Mansfield cirques were made from the U.S. Geological Survey's *Mount Mansfield, Vermont, Quadrangle*, 7.5-minute, 1:24,000-scale topographic map, with a contour interval of 20 feet. This map was compiled from aerial photographs taken in 1947 and 1978, field checked in 1948, photorevised and edited in 1980, and photoinspected in 1983.

Adirondack Mountains, New York

Morphometric measurements for White Brook valley cirque were made from the U.S. Geological Survey's *Wilmington, New York,* 7.5- × 15-minute, 1:25,000-scale metric topographic map, with a contour interval of 10 meters. This map was compiled from aerial photographs taken in 1976, field checked in 1976, and edited in 1978.

Catskill Mountains, New York

Morphometric measurements for Johnson Hollow cirque were made from the U.S. Geological Survey's *Prattsville and Roxbury, New York, Quadrangles*, 7.5-minute, 1:24,000-scale, with a contour interval of 20 feet. These maps were compiled from aerial photographs taken in 1943, and field checked and edited in 1945; the Roxbury quadrangle was also photo-inspected in 1981.

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CIRQUES OF THE PRESIDENTIAL RANGE, NEW HAMPSHIRE, AND SURROUNDING ALPINE AREAS IN THE NORTHEASTERN UNITED STATES

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ABSTRACT Evidence for rejuvenation of cirque glaciers following wastage of continental ice remains elusive for the Presidential Range and Mount Moosilauke of New Hampshire, Mount Katahdin and the Longfellow Mountains of Maine, and the Adirondack Mountains of New York. At Ritterbush Pond in the northern Green Mountains of Vermont, radiocarbon ages from lake sediment cores suggest that a low-altitude valley head, located upvalley of a series of cross-valley moraines, was ice-free by 11,940 ^{14}C yrs BP (Bierman *et al.*, 1997). Although some workers argue that these moraines in Vermont are evidence for cirque glaciation, the moraines could have been formed by a tongue of continental ice during deglaciation. At Johnson Hollow Brook valley in the Catskill Mountains of New York, a radiocarbon age from basal sediments in a pond dammed by a moraine suggests that glacier ice may have persisted until 10,860 ^{14}C yrs BP (Lederer and Rodbell, 1998). Because this moraine appears to have been deposited by a cirque glacier, the radiocarbon age provides the best evidence in the northeastern United States for cirque glaciation post-dating recession of continental ice. Cirque morphometric data, compiled from newly available topographic maps, add to the conundrum that these two sites in the Green and Catskill Mountains should not be nearly as favorable for maintaining local glaciers post-dating icesheet recession as higher-altitude and better-developed cirques in the Presidential Range and Mount Katahdin, where evidence for post-icesheet cirque glaciers is lacking.

RÉSUMÉ *Les cirques du Presidential Range (New Hampshire) et des régions alpines avoisinantes du nord-est des États-Unis.* Les signes de rajeunissement des glaciers de cirque après le retrait de l'inlandsis demeurent ambigus au Presidential Range (New Hampshire), au mont Katahdin et dans les Longfellow Mountains (Maine), ainsi dans les Adirondack (New York). Au Ritterbush Pond, dans le nord des Green Mountains du Vermont, les datations au radiocarbone de sédiments lacustres ont établi que l'extrémité d'une vallée de basse altitude, située en amont d'une série de moraines transversales, était libre de glace vers 11 940 BP. Bien que certains chercheurs croient que ces moraines témoignent d'une glaciation de cirque, il est aussi possible que les moraines aient été mises en place par une langue glaciaire pendant la déglaciation. Dans la vallée du Johnson Hollow Brook dans les Catskill (New York), une datation au radiocarbone de sédiments de base dans un étang fermé par une moraine indique que la glace pourrait avoir persisté jusqu'à vers 10 860 BP (Lederer et Rodbell, 1998). Comme cette moraine semble avoir été mise en place par un glacier de cirque, la datation fournit le meilleur indice de glaciation de cirque survenue après le retrait de l'inlandsis dans le nord-est des États-Unis. Les données sur la morphométrie des cirques, compilées à partir des nouvelles cartes topographiques, creusent le mystère en établissant que les deux sites des Green Mountains et des Catskill seraient moins propices au maintien des glaciers locaux après le retrait de l'inlandsis que le seraient d'autres cirques mieux calibrés et à plus haute altitude du Presidential Range et au mont Katahdin où on ne trouve pas d'indices de cirques glaciaires après le retrait de l'inlandsis.

ZUSAMMENFASSUNG *Alpine Vergletscherung und Kar-Morphometrie in der Presidential Range, New Hampshire, und in den umgebenden bergigen Gebieten in Nordost-U.S.A.* Hinweise auf eine Verjünjung der Kar-Gletscher als Folge des Schwunds von kontinentalem Eis bleiben für die Presidential Range von New Hampshire, den Mount Katahdin und die Longfellow Mountains von Maine sowie die Adirondacks von New York schwer greifbar. Am Ritterbush Pond in den nördlichen Green Mountains von Vermont zeigen Radiokarbondatierungen von Seesedimentkernen, dass ein Talhaupt von niedriger Höhe, das sich oberhalb einer Serie von Quermoränen befand, um etwa 11 940 ^{14}C Jahre v.u.Z. (Bierman *et al.*, 1997) eisfrei war. Obwohl einige Forscher diese Moränen für einen Beweis der Kar-Vergletscherung halten, könnten diese Moränen durch eine Zunge kontinentalen Eises während der Enteisung gebildet worden sein. Im Johnson Hollow Brook-Tal in den Catskill Mountains von New York erlaubt eine Radiokarbondatierung von Basis-Sedimenten in einem von einer Moräne eingedämmten Teich die Annahme, dass Gletscher-Eis bis etwa um 10 860 ^{14}C Jahre v.u.Z. (Lederer und Rodbell, 1998) gedauert hat. Da diese Moräne durch einen Kar-Gletscher abgelagert worden zu sein scheint, liefert die Radiokarbondatierung den besten Beweis für Kar-Vergletscherung nach dem Rückzug des Kontinental-Eises im Nordosten der U.S.A. Morphometrische Kar-Daten, die mittels neu verfügbarer topographischer Karten zusammengestellt wurden, tragen zu dem Rätsel bei, dass nämlich diese zwei Plätze in den Green und Catskill Mountains für die Aufrechterhaltung lokaler Gletscher nach dem Rückzug der Eisdecke bei weitem nicht so geeignet hättensein sollen wie höher gelegene und besser entwickelte Kare in der Presidential Range und am Mount Katahdin, wo Belege für Kar-Gletscher nach dem Rückzug der Eisdecke fehlen.

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INTRODUCTION

The purpose of this report is twofold: 1) to evaluate evidence for local (cirque) glaciation in the northeastern United States (Fig. 1) in light of work since the publication by Waitt and Davis (1988), and 2) to summarize data on cirque morphometry, especially schrund altitudes (Fig. 2), measured on new topographic base maps that have become available in the past 10 years for the Presidential Range in New Hampshire, the Longfellow Mountains and Mount Katahdin in Maine, selected sites in the Green Mountains of Vermont, as well as the Adirondack and Catskill Mountains in New York. Although the base map has been available for over 30 years, complete cirque morphometric data from Mount Moosilauke in New Hampshire are summarized for the first time. Many of the ideas in this paper have been presented during the past decade on society field trips in the Presidential Range (Davis *et al.*, 1988, 1993, 1996a), Mount Katahdin (Davis and Caldwell, 1994), and the Green Mountains (Wright *et al.*, 1997b). Cirque morphometry maps are included in this paper; photographs and geologic maps are not included, although reference is made to the literature where appropriate.

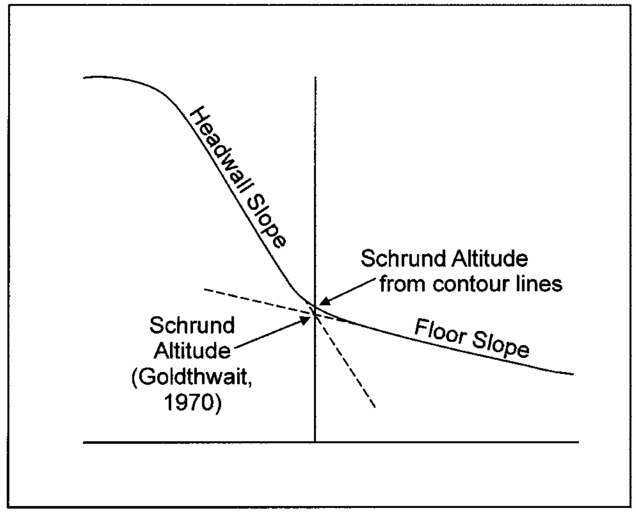


FIGURE 2. Schematic illustration for estimation of schrund altitude (after Goldthwait, 1970, Fig. 9, p.100).

Estimation schématisée de l'altitude d'une rimaye (selon Goldthwait, 1970, fig. 9, p. 100).

RELATIVE AND RADIOMETRIC CIRQUE GLACIER CHRONOLOGIES

PRESIDENTIAL RANGE, NEW HAMPSHIRE

The name Goldthwait is synonymous with not only the glacial history of New Hampshire (Goldthwait *et al.*, 1951) but also with our understanding of cirque glaciation in the Presidential Range (Figs. 1, 3) of the White Mountains (see W.B. Thompson, this volume). James W. Goldthwait was the first to carry out an extensive study of glaciation in the Presidential Range (1913a, b, 1916, 1938), where he reached three major conclusions: 1) the uplands above the cirques were eroded by both fluvial and glacial activity, 2) the cirques were carved by alpine glaciers, as opposed to continental ice, stream erosion, or frost action, and 3) continental glaciation followed the last cirque glacier activity. J.W. Goldthwait's 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. J.W. Goldthwait (1913a) did not support the concept that local glaciers extended far down valleys from an icecap centered on the Presidential Range, as proposed by Packard, Vose, and the Hitchcocks (see W.B. Thompson, this volume).

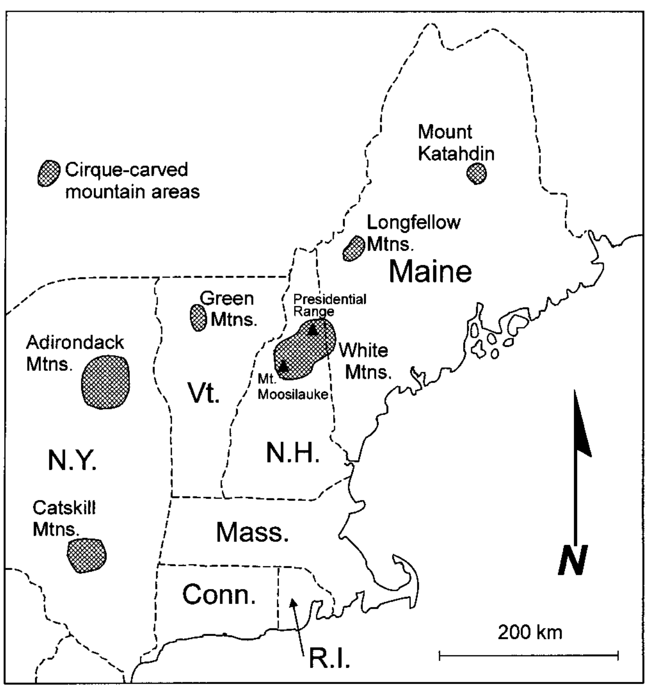


FIGURE 1. Index map for alpine areas with cirques in northeastern United States (after Bradley, 1981, and Ackerly, 1989).

Carte repère du nord-est des États-Unis illustrant les régions alpines où se trouvent des cirques (selon Bradley, 1981 ; Ackerly, 1989).

Over the next two decades, only two workers strongly disputed the conclusions of J.W. Goldthwait concerning the timing of continental and cirque glaciation in the Presidential Range. Johnson (1917, 1933) suggested that 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 and on Mount Katahdin, Maine; however, neither author provided a convincing explanation for the till of farther northern provenance on the cirque floors in the two areas.

Richard P. Goldthwait (1939, 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's in support of cirque glacier

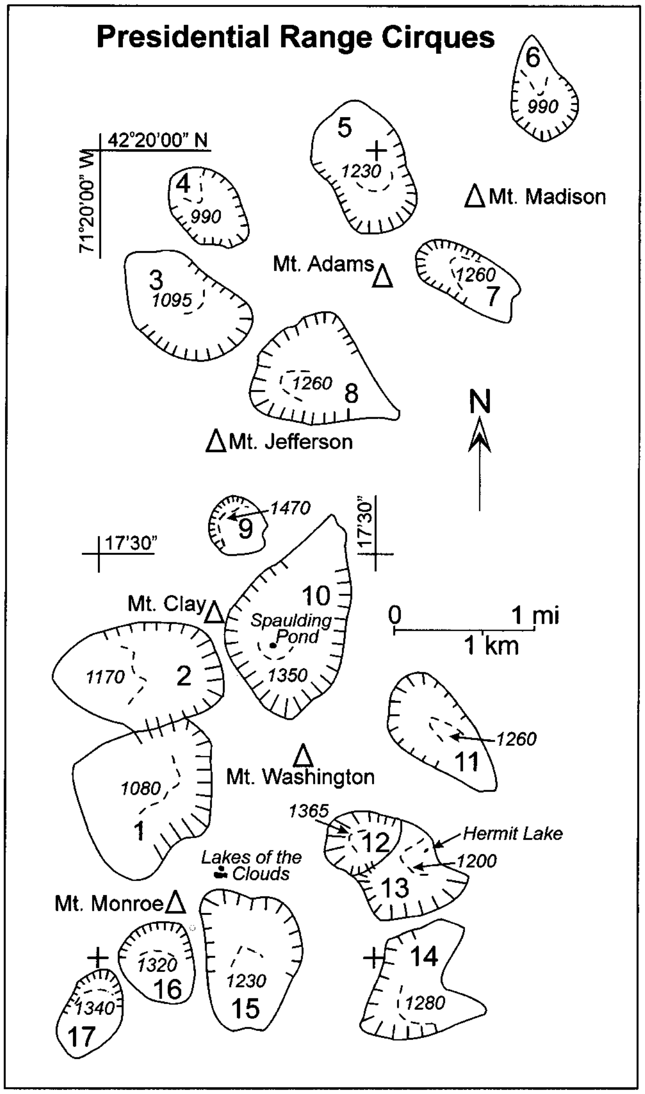


FIGURE 3. Cirque morphometric map for Presidential Range. Solid lines indicate outlines of cirques, hachured solid lines are cirque headwalls, dashed lines are cirque schrunds, with altitudes indicated in meters. Numbers refer to Table I. Small water bodies shown in black for cirques #10 and 13.

Morphométrie des cirques du Presidential Range. Les lignes pleines font le pourtour des cirques, les lignes hachurées montrent les murs de rimaye et les tirets représentent les rimayes (altitude en m). Les chiffres se rapportent au tableau I. Les cirques n^{os} 10 et 13 comprennent chacune une petite étendue d'eau.

activity preceding the last overriding by continental ice, but he also observed roche moutonnées on cirque floors along with striae and grooves on cirque headwalls, which he believed could only have been formed by continental ice. In his 1970 paper, R.P. Goldthwait reviewed his earlier work and provided pebble lithology data for till sites on the uplands and in north-facing cirques, which supported the view that till in the Presidential Range was deposited by continental ice. Also, in this latter paper, R.P. Goldthwait presented morphometric data on cirques and altitudinal estimates of firn lines for the former cirque glaciers in the Presidential Range. 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 1950s, W.F. Thompson (1960a, b, 1961) analyzed aerial photographs of the Presidential Range and Mount Katahdin in Maine and refuted the Goldthwaits' view by arguing that the steep headwalls and sharp arêtes were indicative of active cirque glaciers following continental icesheet deglaciation. W.F. Thompson's primary field contribution was an experiment to test the origin of striae whereby he painted one of the striated portions of the Tuckerman Ravine headwall. Although W.F. Thompson did not present field data to support his view, he believed that moraines of cirque glaciers had been obliterated by post-glacial mass wasting processes. Work in Tuckerman Ravine during the late 1980s by D.J. Thompson (1990, this volume) 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 to the south. However, Gerath and Fowler (1982), Fowler (1984), Gerath *et al.* (1985), Davis and Waitt (1986), and Waitt and Davis (1988) examined the diamicts at the cirque mouths and concluded that the sediments are not till, but rather debris flow deposits. Bradley (1981) also noted fresh grooves across the painted surface on the Tuckerman Ravine headwall, suggesting that if recent snow/ice avalanches could erode bedrock, then perhaps cirque glaciers could also striate cirque headwalls. In 1998, few cobbles and boulders remain on W.F. Thompson's painted surface, and paint is only preserved in the deepest grooves. However, paint also does not survive 30 years on trail signposts exposed to the severe weather conditions in the Presidential Range, so the significance of W.F. Thompson's experiment remains uncertain.

Opportunities for developing a radiocarbon chronology for the deglaciation of cirques in the Presidential Range are limited because of the small number of tarns. Spaulding Pond in the Great Gulf and Hermit Lake in Tuckerman Ravine (Fig. 3), although shallow, may provide useful continuous post-glacial records of sediment accumulation and should be cored. Lakes close by cirques in the Presidential Range have provided minimum radiocarbon ages for ice retreat (Davis *et al.*, 1980; Spear, 1989; Spear *et al.*, 1994; Miller and Spear, this volume).

Organic material from sediments at the base of a core retrieved from Lost Pond at an elevation of 650 m in Pinkham Notch on the east side of the Presidential Range (just off right margin of Fig. 3) provide a radiocarbon age of 12,870 ± 370 yrs BP (QL-985; Spear *et al.*, 1994; all ages reported in this paper are in ^{14}C yrs BP). 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 (Fig. 3) have 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 (Fig. 4), which provides a radiocarbon age of 13,000 ± 400 yrs BP (QL-1133; Spear, 1989). Given the model that continental ice thinned, separated, stagnated, and dissipated over the mountains of northern New England during Late Wisconsinan deglaciation (Goldthwait and Mickelson, 1982; Hughes *et al.*, 1985; Stone and Borns, 1986; Borns, 1987; Davis and Jacobson, 1987; Thompson and 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.

Current work by the author and Paul Bierman at the University of Vermont 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. Included are samples from two large boulders on the tongue-shaped rock glacier on the floor of Tuckerman Ravine to determine the relative age of this cirque deposit (see D.J. Thompson, this volume). Laboratory preparation and analyses of these samples are ongoing.

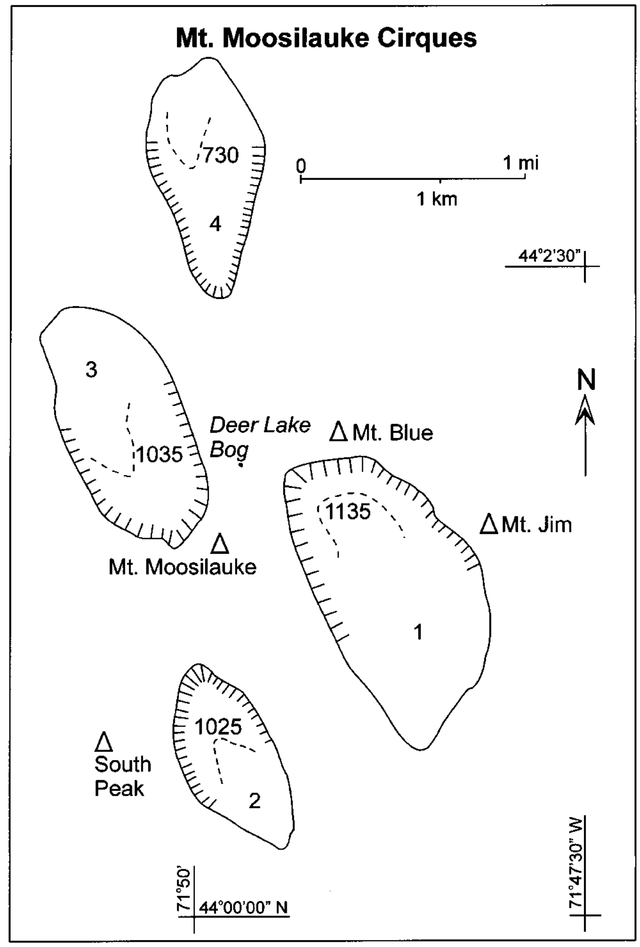


FIGURE 4. Cirque morphometric map for Mount Moosilauke. Solid lines indicate outlines of cirques, hachured solid lines are cirque headwalls, dashed lines are cirque schrunds, with altitudes indicated in meters. Numbers refer to Table I.

Morphométrie des cirques au mont Moosilauke. Les lignes pleines font le pourtour des cirques, les lignes hachurées montrent les murs de rimaye et les tirets représentent les rimayes (altitude en m). Les chiffres se rapportent au tableau I.

MOUNT MOOSILAUKE, NEW HAMPSHIRE

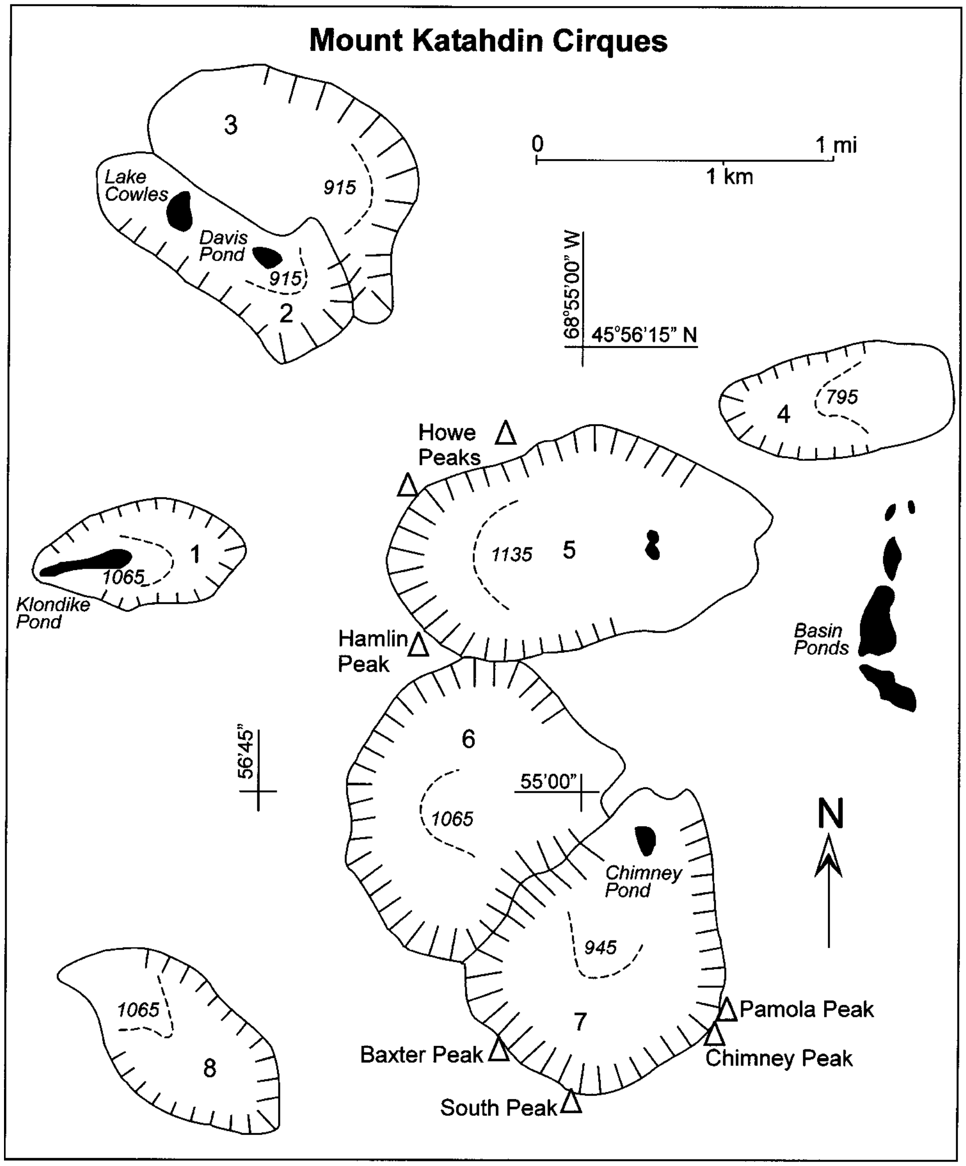
Mount Moosilauke is the second highest massif in the western White Mountains of New Hampshire (Figs. 1, 4). Of all other cirques in the White Mountains outside the Presidential Range, those on Mount Moosilauke are the best-developed. Haselton (1975) described evidence for continental ice overriding the summit areas and noted three cirques on Mount Moosilauke: Jobildunk, Gorge Brook, and Benton Ravines. Although he did not recognize any moraines on cirque floors, Haselton (1975) remained open to the possibility that cirque glaciers post-dated recession of continental ice. Unfortunately, none of the cirques on Mount Moosilauke have tarns that might yield sediment cores for radiocarbon dating.

MOUNT KATAHDIN, MAINE

Nearly all previous researchers at Mount Katahdin (Tarr, 1900; Antevs, 1932; Thompson, 1960a, b, 1961; Caldwell, 1966, 1972, 1998; Caldwell and Hanson, 1982, 1986) have promoted steep headwalls and sharp arêtes as evidence that cirque glaciers post-date continental icesheet deglaciation here (Figs 1, 5); they believe such features could not withstand the effects of an overriding icesheet. These workers also interpreted landforms on cirque floors as moraines and the moraine damming Basin Ponds on the east flank of Mount Katahdin (Fig. 5) to be formed, at least in part, by cirque glaciers. However, Davis (1976, 1978, 1983, 1989) reported observations from Mount Katahdin similar to those made in the Presidential Range by the Goldthwait's as evidence against post-icesheet cirque glacier activity. These observations included: 1) a lack of looped moraines on cirque floors, with the bumps in topography on cirque floors noted by others being hummocky till or landslide deposits rather than moraines, 2) till of a northern provenance on all cirque floors, with especially high percentages of erratic pebble lithologies on the floor of Northwest Basin, a northwest-facing cirque, and 3) roche moutonnées indicating upvalley ice flow on the floor of Northwest Basin. Although striae were not found on cirque headwalls, glacially lpolished surfaces were noted about halfway up Cathedral arête (Davis, 1976) and on Knife Edge arête (Davis *et al.*, 1996b). Along with

FIGURE 5. Cirque morphometric map for Mount Katahdin, Maine. Solid lines indicate outlines of cirques, hachured solid lines are cirque headwalls, dashed lines are cirque schrunds, with altitudes indicated in meters. Cirque numbers refer to Table I. Water bodies shown in black for cirques #1, 2, 5, and 7, and along Basin Ponds moraine east of cirques #5, 6, and 7.

Morphométrie des cirques au mont Katahdin. Les lignes pleines font le pourtour des cirques, les lignes hachurées montrent les murs de rimaye et les tirets représentent les rimayes (altitude en m). La numérotation des cirques est celle du tableau I. Les petites étendues d'eau sont en noirs dans les cirques nos 1, 2, 5 et 7 et le long de la moraine de Basin Pond à l'est des cirques nos 5, 6 et 7.

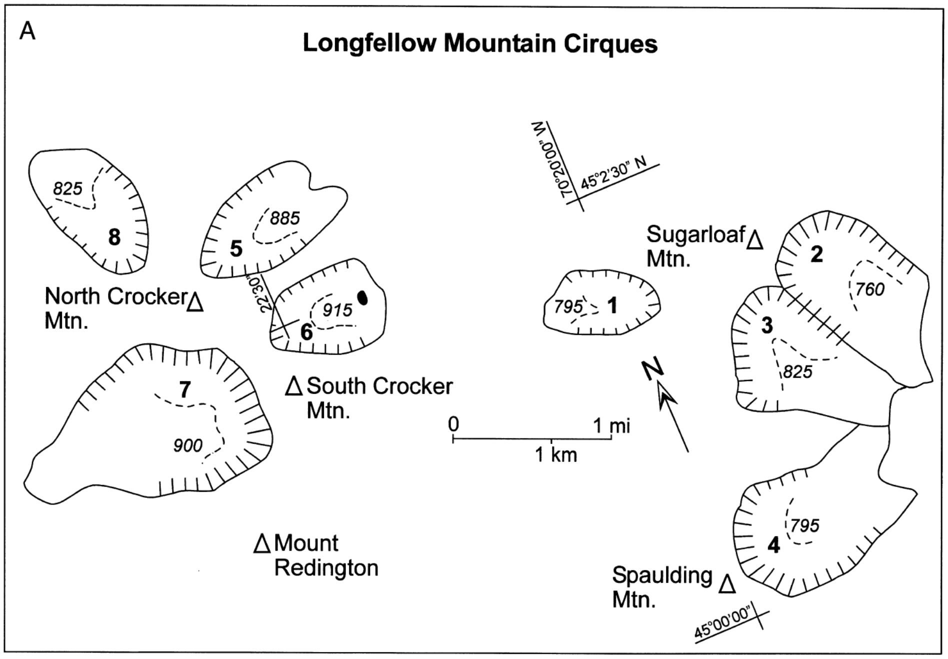


thinness of soils, limited weathering of erratics located near the summit, and theoretical ice profiles, the glacially polished surfaces suggest that Mount Katahdin was overridden by a warm-based continental icesheet at some time during the late Wisconsinan glaciation (Davis, 1989).

Davis (1976, 1978, 1983, and 1989) argued that the Basin Ponds moraine was completely formed by continental ice to the east because: 1) the pebble fraction is 10 to 44% erratic, 2) its morphology is convex westward and follows a contour along the east slope of Mount Katahdin, 3) the moraine extends both north and south beyond the mouths of the three east-facing cirques, 4) there is too little space (< 20 m) between the moraine and Keep Ridge for a cirque glacier, where only a small drainage channel occurs, 5) several smaller arcuate and parallel ridges downslope of the Basin Ponds moraine that lie north and south of the cirque mouths could only be formed by receding continental ice to the east, and 6) along the south slope of Mount Katahdin, where there are no cirques, nearly continuous ridges extend for about 8 km at about the same altitude of the Basin Ponds moraine, so could only be formed by continental ice. Thus, the Basin Ponds moraine and the moraines on the south slope of Mount Katahdin were most likely formed during a late-glacial (nunatak) phase of late Wisconsinan glaciation.

Based on field data from cirques, Davis (1976, 1978) argued that the Late Wisconsinan regional snow-line on Mount Katahdin was too high to support formation of cirque glaciers following icesheet recession, the same case argued by R.P. Goldthwait (1970) for the Presidential Range.

No radiocarbon dating besides that presented in Davis and Davis (1980) is available for lakes and bogs in the Mount Katahdin area. However, preliminary cosmogenic nuclide analyses of ^{10}Be and ^{26}Al from one boulder on a recessional moraine (P. R. Bierman, oral commun., 1998) suggest that the earliest radiocarbon ages from bog and pond basal sediments on the moun-



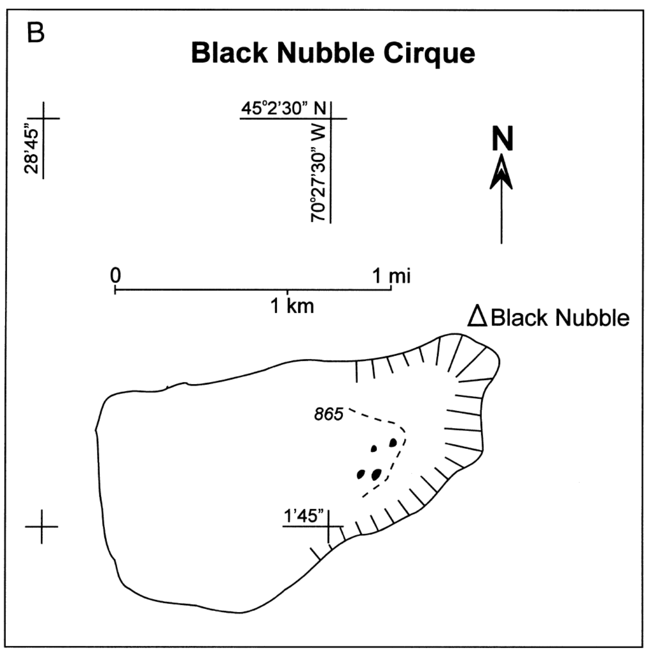


FIGURE 6. A. Cirque morphometric maps for Longfellow Mountains, Maine. Solid lines indicate outlines of cirques, hachured solid lines are cirque headwalls, dashed lines are cirque schrunds, with altitudes indicated in meters. Numbers for cirques on Sugarloaf and Crocker Mountains refer to Table I; only cirque #6 has small water body, shown in black. B. Black Nubble cirque; water bodies shown in black.

A. Morphométrie des cirques des Longfellow Mountains, dans le Maine. Les lignes pleines font le pourtour des cirques, les lignes hachurées montrent les murs de rimaye et les tirets représentent les rimayes (altitude en m). La numérotation des cirques est celle du tableau I ; seul le cirque n^o 6 renferme une petite étendue d'eau (en noir). B. Le cirque Black Nubble ; les étendues d'eau sont en noir.

tain lag ice retreat by several thousand years, as suggested by Davis and Davis (1980). Additional samples of polished bedrock from the Knife Edge arête (Davis *et al*, 1996b), along with boulders on cirque floors, lateral moraines, and the lowlands surrounding Mount Katahdin, are being analyzed for ^{10}Be and ^{26}Al.

LONGFELLOW MOUNTAINS, MAINE

In their study of glaciation in west-central Maine, Borns and Calkin (1977) concluded that ten cirque-like basins in the Longfellow Mountains (Figs. 1, 6) showed no evidence for reactivation by local ice during or subsequent to dissipation of Late Wisconsinan continental ice. Deposits and landforms related to these basins could be explained by continental glaciation and subsequent stagnation of this ice. No looped moraines or similar deposits were found in these cirques. Furthermore, examination of clast content in till forming the floor of two of the best-formed cirques, these facing eastward on Crocker Mountain (Fig. 6a), revealed about 50% erratic lithologies. Based on such data, Borns and Calkin (1977) concluded that the regional snow-line rose to an altitude above the Longfellow Mountains prior or during their emergence from the receding continental icesheet.

Borns and Calkin (1977) reported radiocarbon ages from the lowlands in the Longfellow Mountains between 10,030 ± 180 yrs BP (CY-2464) and 10,860 ± 160 yrs BP

(M. Stuiver, written commun., 1970). These ages are supported by more recently obtained radiocarbon ages for basal sediments from Boundary Pond (11,200 ± 200 yrs BP, GSC-1248, Shilts, 1981) and from Lower Black Pond (11,500 ± 50 yrs BP, OS-7123, Thompson *et al.*, 1996). The latter is an AMS radiocarbon age on terrestrial macrofossils; the former is a conventional radiocarbon age on bulk sediment. Small ponds in two cirques on Crocker Mountain (Fig. 6a) have not been investigated for sediment coring potential. All of these radiocarbon ages are consistent with recent chronological work on glacial Lake Hitchcock (Ridge and Larsen, 1990; Ridge *et al.*, 1996), which led to a new deglaciation model for northern New England proposed by Ridge *et al.* (this volume), who suggest that continental ice remained up to 1500 ^{14}C yrs later than suggested by existing chronologies. However, other radiocarbon ages for deglaciation of the White Mountains support a model for earlier recession of continental ice (Thompson *et al.*, this volume).

GREEN MOUNTAINS, VERMONT

Stewart (1961, 1971) and Stewart and MacClintock (1969) interpreted all drift in the Green Mountains of northern Vermont as the product of a continental icesheet. However, Wagner (1970, 1971) and Connally (1971) proposed that local mountain glaciers post-dated icesheet recession (Figs. 1, 7, 8). These latter authors argued that some low-elevation valley heads are cirques, some lakes are tarns, some ridges on valley floors are moraines built by cirque glaciers, and some deltas down-valley of the ridges were formed by meltwater of cirque glaciers. Waitt and Davis (1988) questioned the conclusions of Wagner (1970, 1971) and Connally (1971), noting that the valley heads do not have the typical bowl shape of cirques, the lakes are too large to be tarns, some of the ridges do not loop across basin floors typical of end moraines formed by cirque glaciers, the deltas are similar to many other ice-contact deposits in the area unrelated to valley heads and therefore are better explained by meltwater from continental ice, and the regional snow-line was too high during the Late Wisconsinan glaciation to support local glaciers at the valley heads. Waitt and Davis (1988) did identify five to seven high-elevation cirques on Mount Mansfield that are comparable to cirques elsewhere in the northeastern United States. Thus, Waitt and Davis (1988) concluded that all of the glacial landforms described by Wagner (1970, 1971) and Connally (1971) were more readily explained by tongues of continental ice rather than by cirque glaciers.

Stephen Wright and his students from the University of Vermont (Wright *et al.*,1997a, b; Loso et al, 1998) re-examined the ridges in the Miller Brook valley (Fig. 7) and concluded that upper valley ridges that were interpreted to be moraines by Wagner (1970) are part of an esker that extends the full length of the valley, ending in a subaqueous fan deposit (Wagner's delta). Wright *et al.* (1997a, b) and Loso *et al.* (1998) also suggested that summer temperatures must drop by about 14°C to support a cirque glacier in Miller Brook valley.

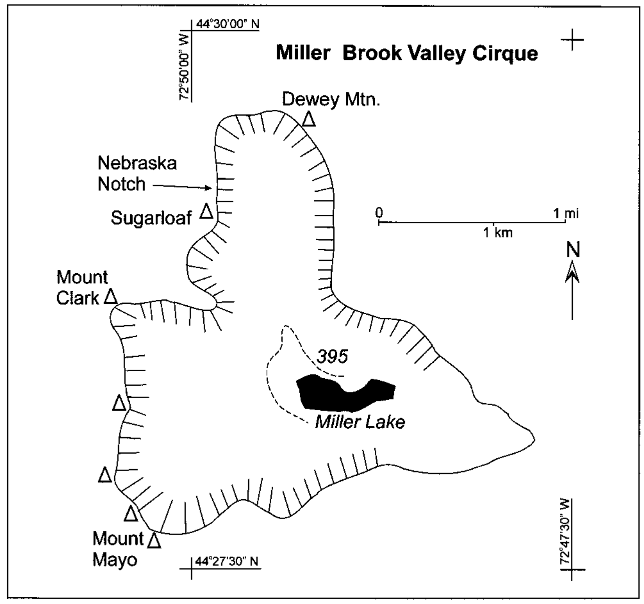


FIGURE 7. Cirque morphometric map for Miller Brook valley cirque, Green Mountains, Vermont. Solid line indicates outline of cirque, hachured solid line is cirque headwall, dashed line is cirque schrund, with altitude indicated in meters. Miller Lake is labeled.

Morphométrie du cirque de la vallée du Miller Brook, dans les Green Mountains (Vermont). Les lignes pleines font le pourtour du cirque, les lignes hachurées montrent les murs de rimaye et les tirets représentent les rimayes (altitude en m).

Sperling *et al.* (1989) interpreted radiocarbon ages obtained from sediment cores in an ephemeral pond alongside the moraine-like feature (esker) in the Miller Brook valley to support the conclusions of Wagner (1970, 1971) and Connally (1971) that cirque glaciers post-dated icesheet deglaciation in the northern Green Mountains. They radiocarbon dated the 275-285 cm interval of one of the sediment cores in Miller Brook valley at 9,280 ± 235 yrs BP (QC1273A), which is considerably younger than other bog- and pond-bottom radiocarbon ages in northern New England. However, Sperling *et al.* (1989) made no mention of the material that they dated or whether gray inorganic silt typical of the late-glacial parts of lake sediment cores was recovered, and thus it is likely that this radiocarbon age underestimates the time of deglaciation.

Wagner (1970) also mapped a series of cross-valley moraines in Ritterbush valley and suggested that the upper part of the valley was a site for cirque glaciation following recession of continental ice (Fig. 8). Sperling *et al.* (1989) recovered lake sediments from a core through a post-glacial delta at the west end of Ritterbush Pond. They radiocarbon-dated the 840-850 cm interval at 10,730 ± 200 yrs BP (QC1272A) and the 850-860 cm interval at 10,090 ± 230 yrs BP (QC1272B). Sperling *et al.* (1989) interpreted the older age to be the more accurate for deglaciation of the valley and suggested that cirque glaciers were present in the valley until about 11,000 yrs ago. However, their pollen analysis of the radiocarbon-dated basal sediments indicates that tundra and spruce-fir pollen zones typical of basal zones of most pollen diagrams in the New England area are not present, suggesting that hundreds if not thousands of years of the post-glacial sediment record may be missing. Again, Sperling *et al.* (1989) did not note whether gray inorganic silt was recovered from the basal part of their Ritterbush Pond sediment core.

In an effort to clarify the timing of ice retreat from the Green Mountains, the author obtained two overlapping sediment cores from the middle of Ritterbush Pond (317 m), with Paul Bierman, Andrea Lini, and their students at the University of Vermont (Lini *et al.*, 1995; Lin *et al.*, 1995; Bierman *et al.*, 1997). Bulk sediments from 569 cm below the mud-water interface (in Ritterbush Core 2) were AMS radiocarbon-dated at 21,860 ± 370 yrs BP (CAMS 20197, $\delta^{13}C = -24‰$). The 479 cm depth of this core was AMS radiocarbon-dated 11,940 ± 90 yrs BP (CAMS 20902; corrected for $\delta^{13}C = -34‰$). Lini *et al.* (1995) argue that the $\delta^{13}C$ values of total organic carbon do not indicate terrestrial vegetation as a major component of primary productivity until about 12,000 ^{14}C yrs ago, thus the 21,860 yrs BP radiocarbon age could reflect a different source of older carbon than the younger ages. From the basal parts of lake sediment cores elsewhere in New England, similarly old radiocarbon ages to the 21,860 ± 370 yrs BP age from Ritterbush Pond have been noted (Davis and Davis, 1980; Davis *et al.*, 1995b; see Ridge *et al.*, this volume). In all cases, these old radiocarbon ages are from sediments with very low values of total organic carbon as determined by loss-on-ignition analyses (Lini *et al.*, 1995; Bierman *et al.*, 1997; P.T. Davis, unpublished data). Five high-altitude cirques on Mount Mansfield (Fig. 9) do not include tarns or other suitable sites for sediment coring and radiocarbon dating.

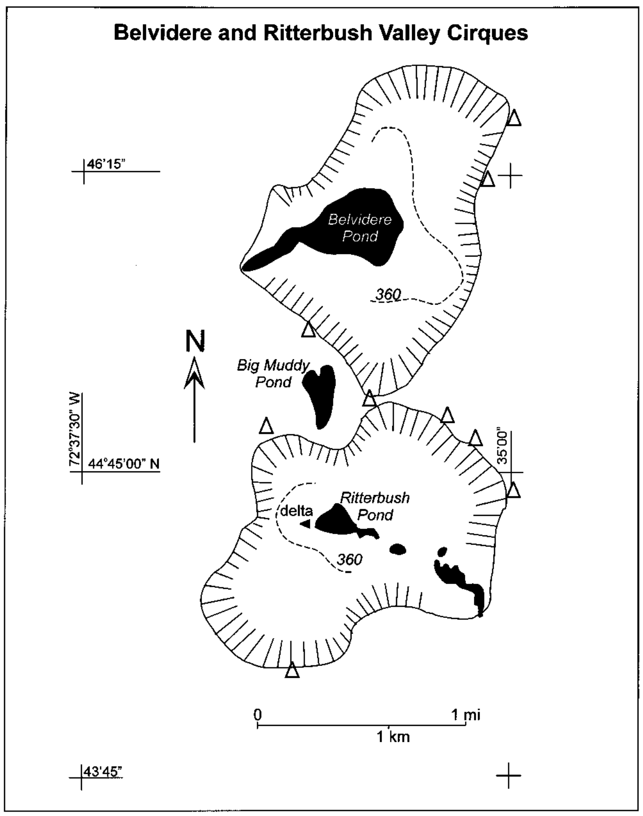


FIGURE 8. Cirque morphometric map for Belvidere (north) and Ritterbush valley (south) cirques, Green Mountains, Vermont. Solid lines indicate outlines of cirques, hachured solid lines are cirque headwalls, dashed lines are cirque schrunds, with altitudes indicated in meters. Water bodies shown in black, except for Belvidere and Ritterbush ponds, which are labeled.

Morphométrie des cirques Belvidere (au nord) et de la vallée du Ritterbush (au sud), dans les Green Mountains (Vermont). Les lignes pleines font le pourtour des cirques, les lignes hachurées montrent les murs de rimaye et les tirets représentent les rimayes (altitude en m). Les étendues d'eau sont en noir.

We also obtained sediment cores from Sterling Pond (917 m), about 60 km south of Ritterbush Pond, in order to compare pollen records from sites at different altitudes (Lin *et al.*, 1995). Bulk sediments from the 522 cm depth near the base of one sediment core from Sterling Pond were radiocarbon-dated at 12,760 ± 70 yrs BP (CAMS 17895). Taken at face value, the radiocarbon ages from Ritterbush and Sterling Ponds in the northern Green Mountains are remarkably similar to those from comparable altitudes in the White Mountains and suggest that continental ice thinned and/or back-wasted rapidly during Late Wisconsinan deglaciation. Palynological analyses from Ritterbush and Sterling Ponds are in progress to determine whether pollen indicators might be useful as a chronological tool where radiocarbon ages are suspect (Lin *et al.*, 1995; Davis *et al.*, unpublished data).

ADIRONDACK MOUNTAINS, NEW YORK

Craft (1976, 1979) examined numerous sites in the High Peaks region of the Adirondack Mountains (Fig. 1) that he suggested were locations of cirque glacier activity during and following disintegration of continental ice. Craft's evidence included over-deepened basins, steep headwalls and side walls, and moraines on basin floors. However, Barclay (1993) re-examined three of the main sites described by Craft (1976) and concluded that the evidence for local glaciers post-dating retreat of continental ice was weak. The Lost Pond depression adjacent to Weston Mountain was found to be lacking key morphometric elements of a cirque, and mapping of glacial lake sediments in the East Roaring Brook valley on the east side of Giant Mountain placed severe constraints on the extent of any local glacier in this valley, assuming that any post-Wisconsinan local glacier existed here at all (Barclay, 1993).

White Brook valley on the northeast side of Whiteface Mountain (Fig. 10) has long been considered to have hosted a local glacier following wastage of continental ice (Alling, 1916, 1919; Johnson, 1917; Craft, 1976, 1979). However, Barclay (1993) showed that an unvegetated bank interpreted by these previous workers as a moraine deposited by a cirque glacier is actually an erosional feature, cut by White Brook as it incised into valley fill deposits. The lithology of pebbles (Craft, 1976; Barclay, 1993) and light mineral fractions (Craft, 1976) of tills in the valley suggest deposition from continental ice flowing southwest, up White Brook valley. Furthermore, the interpretation of a till down-valley from

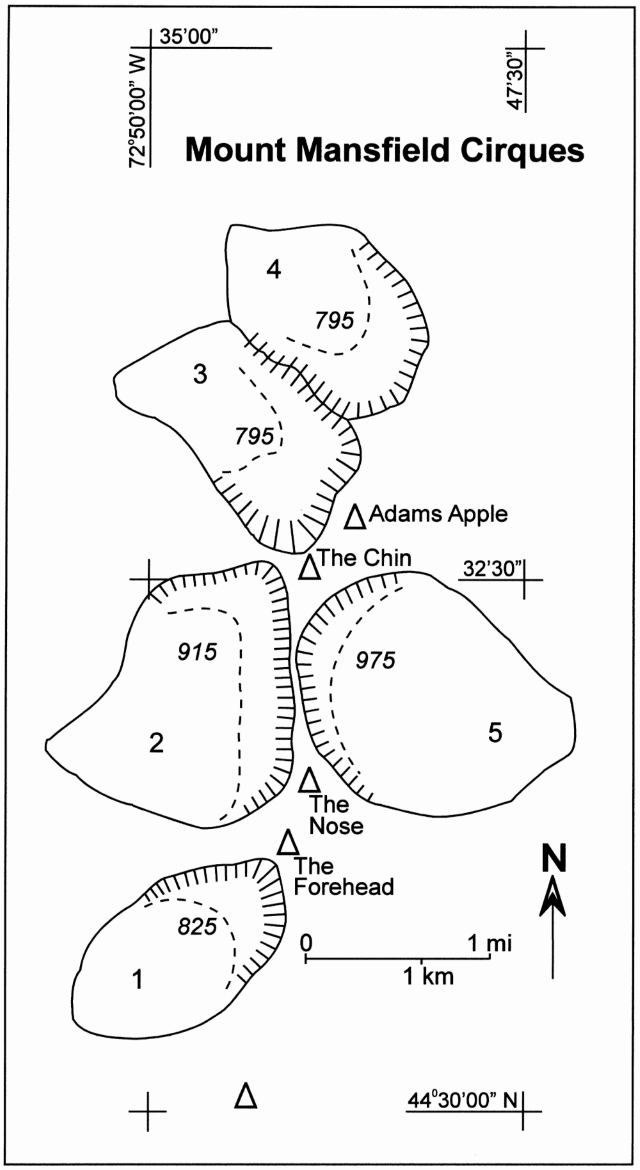


FIGURE 9. Cirque morphometric map for Mount Mansfield, Vermont. Solid lines indicate outlines of cirques, hachured solid lines are cirque headwalls, dashed lines are cirque schrunds, with altitudes indicated in meters. Cirque numbers refer to Table I.

Morphométrie du cirque de vallée du mont Mansfield, au Vermont. Les lignes pleines font le pourtour des cirques, les lignes hachurées montrent les murs de rimaye, les tirets représentent les rimayes (altitude en m). La numérotation des cirques est celle du tableau I.

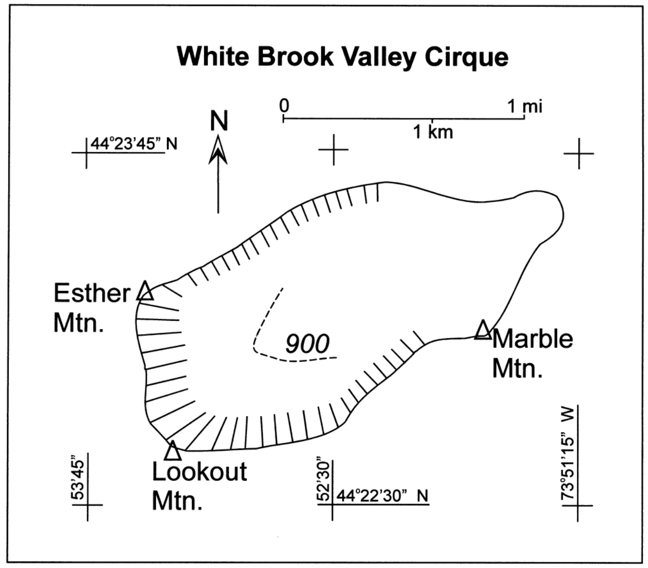


FIGURE 10. Cirque morphometric map for White Brook valley, Adirondack Mountains, New York. Solid line indicates outline of cirque, hachured solid line is cirque headwall, dashed line is cirque schrund, with altitude indicated in meters.

Morphométrie du cirque de la vallée du White Brook, dans les Adirondack (New York). Les lignes pleines font le pourtour des cirques, les lignes hachurées montrent les murs de rimaye et les tirets représentent les rimayes (altitude en m).

the mouth of White Brook valley as a local glacier deposit by Craft (1976) contradicts reconstructions by Franzi (1992) of glacial lake levels in the adjacent Ausable River valley.

Although sediment cores have not been recovered from high-elevation basins in the High Peaks region for radiocarbon dating deglaciation, a sediment core was obtained from Readway Pond at an elevation of 424 m on an outwash plain on the northwestern flank of the Adirondacks. Although this site is far away from any Adirondack cirques, bulk detrital organic material from the 754-764 cm interval of a sediment core yielded a conventional radiocarbon age of 12,640 ± 430 yrs BP (GX-14486), thus providing a minimum-limiting age for continental ice retreat from the Star Lake moraine (Pair and Rodrigues, 1993; Davis *et al.*, 1995a). If the deglaciation model described for the White Mountains in New Hampshire, as well as the Longfellow Mountains and Mount Katahdin in Maine, is also appropriate for deglaciation of the Adirondacks, then cirques in the High Peaks region may have been free of ice as early of 13,000 years ago.

CATSKILL MOUNTAINS, NEW YORK

The Catskill Mountains (Fig. 1) may hold the best evidence for local mountain glaciers following wastage of continental ice. Rich (1906, 1935) identified numerous basins in the Catskills with steep headwalls and side walls, broad floors, and looped end moraines composed of locally-derived till. At the head of Johnson Hollow Brook valley (Schoendorf cirque; Fig. 11), Rich (1906) also identified striae that he interpreted to be formed by a local glacier flowing northeast. Johnson (1917) agreed with all of Rich's (1906) observations except for the striae in Johnson Hollow Brook valley, which he believed were formed by regional flow of continental ice towards the southwest, up the valley. Rich (1935) believed that post-glacial talus deposits caused the cirque headwalls and side walls to be less steep than they were in the past. Cadwell (1986), who concurred with earlier suggestions that a local glacier occupied Schoendorf cirque, suggested that local ice may have persisted long enough to develop a set of glacial terraces on the cirque headwall. Cadwell (1986) also noted a pond

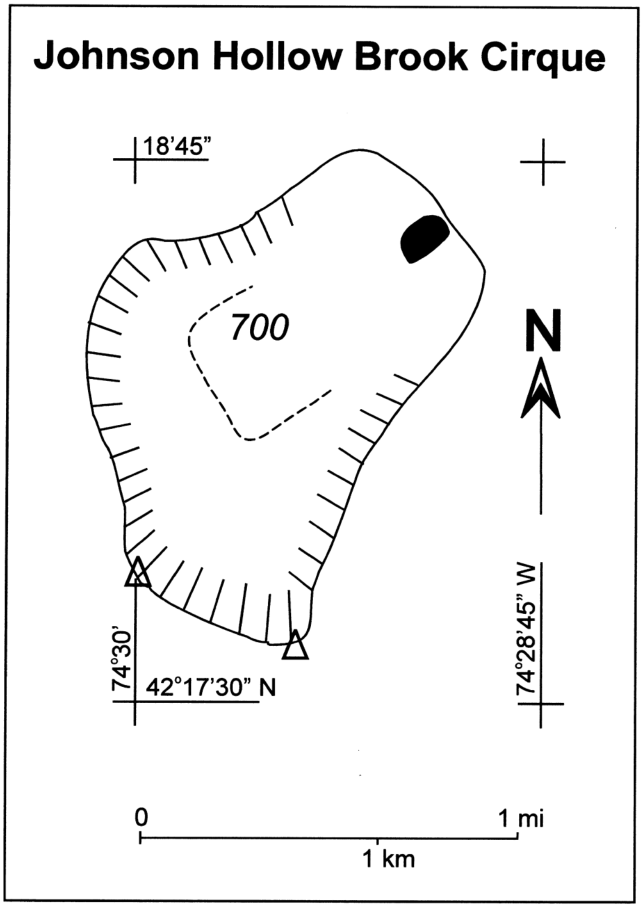


FIGURE 11. Cirque morphometric map for Johnson Hollow Brook, Catskill Mountains, New York. Solid line indicates outline of cirque, hachured solid line is cirque headwall, dashed line is cirque schrund, with altitude indicated in meters. Water body shown in black.

Morphométrie des cirques du Johnson Hollow Brook, dans les Catskill (New York). Les lignes pleines font le pourtour des cirques, les lignes hachurées montrent les murs de rimaye et les tirets représentent les rimayes (altitude en m). L'étendue d'eau est en noir.

dammed by the moraine across the mouth of the cirque, and he suggested that regional deglaciation of continental ice in the Catskills occurred about 15,000 yrs BP.

The moraine-dammed pond in Schoendorf cirque (Fig. 11) attracted the attention of Donald Rodbell and his students at Union College in Schenectady in the middle 1990s. Lederer (1998) and Lederer and Rodbell (1998) confirmed the concave up-valley form of the moraine, determined the moraine's composition to be 98% locally derived clasts, obtained a sediment core from the pond, and conventionally radiocarbon-dated bulk detrital organic material at the 545 cm depth. This depth marked an abrupt transition from pink clay below to organic-rich gyttja above, which was radiocarbon-dated to 10,860 ± 115 yrs BP (GX-23836). If the pink clay is a late-glacial diamict of local origin, as suggested by Lederer (1998), Johnson Hollow Brook valley provides the only radiocarbon age from a cirque in the northeastern United States that suggests persistence of local alpine ice following recession of the last continental icesheet.

THEORETICAL GLACIER PROFILE RECONSTRUCTIONS

Shreve (1985a, b) constructed a theoretical continental ice-surface profile based on esker data to suggest that Mount Katahdin was a nunatak, with ice at the late Wisconsinan maximum reaching only to about the 1100-m altitude on the mountainside. However, field data, such as unweathered erratics in till near the summit areas, and theoretical ice-surface profiles based on Nye's (1952) shear-stress equation

$$\tau_b = \rho g h \sin\alpha \qquad (1)$$

where τ_b is basal shear stress of the ice, ρ is ice density, g is gravity, h is ice thickness, and α is the surface gradient of the ice, suggest that Mount Katahdin was covered by continental ice during the Late Wisconsinan (Davis, 1989). Ackerly (1989) constructed ice-surface profiles for 37 proposed mountain glaciers in northeastern United States using the above shear-stress equation, modified by introducing a shape factor (Nye, 1952) based on the shape of the valley cross section and the ratio of glacier width to depth, such that

$$\tau_b = \rho g h F \sin\alpha \quad \text{and} \quad F > 1 \qquad (2)$$

Given reasonable assumptions for assigned values τ_b and $\sin\alpha$, Ackerly's (1989) reconstructions supported the existence of local alpine glaciers at some time in the past for the Presidential Range, Mount Moosilauke, Mount Katahdin, the Longfellow Mountains, the Catskills, and some sites in the Adirondacks, whether or not these sites were occupied by cirque glaciers subsequent to wastage of continental ice. However, for Wagner's (1970, 1971) proposed sites for cirque glaciers in the Green Mountains, Ackerly (1989) found that reconstructed local ice thickness exceeded the depth of their respective valleys or the altitude of their up-valley cols. He noted that these latter sites were either very shallow basins or very gently sloping valleys. Thus, Ackerly's (1989) theoretical ice-surface profiles support the conclusions of Waitt and Davis (1988), which questioned whether Miller Brook, Ritterbush, and Belvidere valleys were true cirques. Ackerly (1989) did not model ice profiles for the high-altitude cirques on Mount Mansfield (Fig. 9).

CIRQUE MORPHOMETRY

BACKGROUND

Lewis (1938) defined four important characteristics of cirques: 1) steep and usually shattered headwalls and side walls, 2) a gentle rock floor usually with evidence of overdeepening and smoothing, 3) a rock lip or threshold at the mouth, and 4) a rock node at the junction of the headwall and cirque floor. In areas of the world where cirques have undergone substantial post-glacial mass wasting and were last exposed to continental icesheets rather than local glaciers, the latter three characteristics are commonly buried, and therefore difficult to observe, both in the field and even on topographic maps. Evans and Cox (1974, 1995) provide a more quantitative basis for describing cirque morphometry on topographic maps, some of which is adopted in this report.

Although cirques are not as numerous in the northeastern United States as in other mountainous areas of the world (*e.g.*, Andrews, 1965; Andrews and Dugdale, 1971; Williams, 1975; Graf, 1976; Evans, 1977; Evans and Cox, 1995), their altitude, orientation, and form can tell us much about paleoclimatic conditions. Flint (1971, p. 67-70, 133-138) suggested that cirque floor altitudes approximate the orographic snow-line at the time of cirque glacier erosion. In a comparison of six methods for estimating equilibrium-line altitudes (ELAs), Meierding (1982) found that measuring cirque floor altitudes was one of the more rapid methods, albeit one of the more subjective, and was prone to underestimation of ELAs.

Goldthwait (1970) estimated the schrund altitude for former mountain glaciers in the Presidential Range as the elevation at which projections of the steepest headwall long slope and the average cirque floor slope intersect (Fig. 2). In all but two cases for the Presidential Range (Oakes Gulf and Castle Ravine; Goldthwait, 1970, p. 100, Fig. 9), the schrund altitudes determined by this method and a simple measurement of change in spacing between contour lines on topographic maps differed by less than 20 m. Because schrund altitudes are higher than cirque floors, their measurement is a more conservative method for estimating the former depression of ELA than is measuring cirque floor altitudes. Goldthwait's (1970) calculation of a mean summer temperature lowering between 5 and 10°C necessary to support cirque glaciers in the Presidential Range today might be considered a minimum estimate. However, other factors may also be important for estimating paleo-ELAs, such as the amount of winter precipitation or the aspect of cirques, as noted by Goldthwait (1970).

Besides schrund altitude and cirque aspect, other parameters of cirque morphology are also useful for estimating paleoenvironmental conditions of cirque glaciers. For example, cirque length-to-height ratios allow one to estimate glacier surface profiles, which, in turn, could influence erosion potential of cirque glaciers (Embleton and King, 1975, p. 209-210). Cirque length-to-height and length-to-width ratios also allow one to compare degree of cirque development, which can be used to categorize cirques by grade (Evans and Cox, 1995). Cirque volumes can be calculated by: (length $\times$ height $\times$ width/2) (Andrews, 1975), which allows estimation of duration of cirque glaciation if given empirically-derived cirque erosion rates (Andrews, 1972; Andrews and LeMasurier, 1973; Reheis, 1975; Anderson, 1978). Headwall-to-floor slope ratios allow one to estimate the amount of over-deepening of cirques, although other morphometric methods are probably more accurate (Haynes, 1968).

METHODS

Morphometric measurements for cirques in seven different areas of the northeastern United States as measured from topographic maps, most published since 1988, are summarized in Table I. Specific names, scales, and contour intervals for these maps are provided in the Appendix 1. Cirque grade, a qualitative evaluation of cirque form, follows the classification of Evans and Cox (1995). Aspect is measured as the direction faced by the central headwall, with the headwall orientation defined as a perpendicular to the long axis of each cirque; negative values increase from 360° to 180° to facilitate averaging. Schrund altitudes are measured as the most obvious break in slope between cirque headwalls and floors as seen by a change in spacing between contour lines (Fig. 2); these provide slightly higher estimates than the method of Goldthwait (1970). Numerous measurements of slope between tops of headwalls and schrund altitudes provide average headwall slopes somewhat lower than values for steepest long slopes of headwalls as measured by Goldthwait (1970). Average floor inclinations below schrund altitudes are measured in similar fashion to Goldthwait (1970).

PRESIDENTIAL RANGE, NEW HAMPSHIRE

Four basins in the Presidential Range in addition to those summarized by Goldthwait (1970) have been identified as cirques in this report: Ammonoosuc, Burt, and Cascades Ravines and "Franklin Basin" (Table I, Figs. 1, 3). These four cirques, as well as "Sphinx Basin" and "Monroe Basin" recognized by Goldthwait (1970), are all ranked grade 4, or poor in the classification of Evans and Cox (1995). Although there may be some doubt whether these basins should be classified as cirques, well-developed characteristics (namely steep headwalls and side walls) compensate for weak ones (namely lack of broad and/or gently sloping floors). Only four cirques in the Presidential Range ranked grade 1 (Upper Great Gulf, Huntington, Upper Tuckerman, and Lower Tuckerman Ravines); King Ravine and Upper Oakes Gulf were ranked only grade 2 because of their steeply sloping floors and weakly developed side walls, respectively. Goldthwait (1970) suggested that snow drifted by prevailing southwesterly winds during glacial periods was important in determining locations of cirques in the Presidential Range; this suggestion is supported by the northern and eastern aspects of the six cirques graded 1 and 2. The mean aspect for all Presidential Range cirques is about 53° azimuth, with a standard deviation of 88° (Table I). The average schrund altitude of 1270 m for the six cirques graded 1 and 2 is about 40 m higher than the average schrund altitude of 1230 m for all Presidential Range cirques (Table I). This difference could indicate greater duration of cirque glacier erosion in basins with higher altitudes or more recent occupation by cirque glaciers. However, the schrund altitudes labeled "RPG" in Table I (from Goldthwait, 1970, Table 1, p. 89) suggest grade 1 and 2 cirques are only about 25 m higher than the average schrund altitude for all cirques, about 1245 m.

From morphometric data on height, width, and length of cirques (Table I), length-to-height (L:H) and length-to-width (L:W) ratios and cirque volumes are calculated. The mean

TABLE I

Cirque morphometric characteristics, New England area, U.S.A.

Cirque Name	Cirque Grade[1]	Cirque Aspect (°)[2]	Schrund Altitude (m)[3]	Schrund Altitude (m)[3]	Aver. Height (m)[4]	Aver. Width (m)[5]	Cirque Length (m)[6]	Average headwall slope (°)[7]	Aver. floor inclination (°)[8]	Length: Height Ratio	Length: Width Ratio	Headwall: Floor Slope Ratio	Cirque Volume (km^3)[9]
Presidential Range, N.H.				**RPG**									
1. Ammonoosuc Ravine	4	–80	1080		530	1370	1525	30	10	2.88	1.11	3.00	0.55
2. Burt Ravine	4	–95	1170		630	990	1830	27	13	2.90	1.85	2.08	0.57
3. Castle Ravine	3	–45	1095	1067	585	840	1525	30	13	2.61	1.82	2.31	0.37
4. Cascade Ravine	4	–45	990		435	610	1065	29	11	2.45	1.75	2.64	0.14
5. King Ravine	2	–20	1230	1165	660	915	1525	36	17	2.31	1.67	2.12	0.46
6. Bumpus Basin	3	5	990	964	360	535	1065	27	10	2.96	1.99	2.70	0.10
7. Madison Gulf	3	120	1260	1214	420	535	1145	32	13	2.73	2.14	2.46	0.13
8. Jefferson Ravine	3	105	1260	1262	570	990	1675	38	11	2.94	1.69	3.45	0.47
9. "Sphinx Basin"	4	100	1470	1409	270	535	610	31	21	2.26	1.14	1.48	0.04
10. Upper Great Gulf	1	10	1350	1342	650	1065	2135	42	9	3.28	2.00	4.67	0.74
11. Huntington Ravine	1	135	1260	1287	630	610	1525	43	11	2.42	2.50	3.91	0.29
12. Upper Tuckerman Ravine	1	110	1365	1378	285	535	610	38	11	2.14	1.14	3.45	0.05
13. Lower Tuckerman Ravine	1	50	1200	1232	435	685	915	27	6	2.10	1.34	4.50	0.14
14. Gulf of Slides	3	65	1280	1262	340	1065	990	30	12	2.91	0.93	2.50	0.18
15. Upper Oakes Gulf	2	160	1230	1275	385	915	1415	32	3	3.68	1.55	10.67	0.25
16. "Monroe Basin"	4	165	1320	1342	310	685	915	35	9	2.95	1.34	3.89	0.10
17. "Franklin Basin"	4	160	1340		340	455	990	33	17	2.91	2.18	1.94	0.08
average	**2.8**	**52.9**	**1228.8**	**1246.1**	**452.5**	**751.4**	**1222.5**	**33.6**	**11.2**	**2.70**	**1.64**	**3.62**	**0.25**
standard deviation	**1.2**	**87.8**	**131.3**	**124.1**	**137.6**	**257.4**	**424.7**	**5.0**	**4.2**	**0.42**	**0.44**	**2.08**	**0.21**
												total volume	**4.21**
Mt. Moosilauke, N.H.				**GMH**									
1. Jobildunk Ravine	2	140	1135	1130	260	950	1950	23	5	7.50	2.05	4.60	0.24
2. Gorge Brook Ravine	3	150	1025	1220	305	500	1440	22	7	4.72	2.88	3.14	0.11
3. Benton Ravine	3	–30	1035	1100	610	750	1780	32	12	2.92	2.37	2.67	0.41
4. Little Tunnel Brook Ravine	3	–10	730		425	600	1550	23	9	3.65	2.58	2.56	0.20
average	**2.8**	**62.5**	**981.3**	**1150.0**	**400.0**	**700.0**	**1680.0**	**25.0**	**8.3**	**4.70**	**2.47**	**3.24**	**0.24**
standard deviation	**0.5**	**95.7**	**174.7**	**62.4**	**156.4**	**195.8**	**229.1**	**4.7**	**3.0**	**2.01**	**0.35**	**0.94**	**0.12**
												total volume	**0.96**
Mount Katahdin, Me.				**PTD**									
1. Klondike Basin	1	–90	1065	1052	215	455	1145	26	2	5.33	2.52	13.00	0.06
2. Upper Northwest Basin	1	–45	915	899	275	535	1600	38	8	5.82	2.99	4.75	0.12
3. Lower Northwest Basin	2	–45	915		365	760	1370	39	9	3.75	1.80	4.33	0.19
4. Little North Basin	2	90	795	808	305	535	1220	25	5	4.00	2.28	5.00	0.10
5. North Basin	1	90	1135	1067	510	990	1980	39	8	3.88	2.00	4.88	0.50
6. Great Basin	2	85	1065	1052	440	1450	1525	39	8	3.47	1.05	4.88	0.49
7. South Basin	1	15	945	945	650	1065	1675	47	4	2.58	1.57	11.75	0.58
8. Witherle Ravine	3	–50	1065	1143	475	535	1450	21	15	3.05	2.71	1.40	0.18
average	**1.6**	**6.3**	**987.5**	**995.1**	**404.4**	**790.6**	**1495.6**	**34.3**	**7.4**	**3.98**	**2.12**	**6.25**	**0.28**
standard deviation	**0.7**	**73.7**	**112.9**	**115.7**	**142.3**	**350.4**	**266.0**	**9.1**	**3.9**	**1.09**	**0.64**	**3.97**	**0.21**
												total volume	**2.21**

TABLE I (continued)

Cirque morphometric characteristics, New England area, U.S.A.

Longfellow Mountains, Me.				**HWB/PEC**									
1. Sugarloaf Mountain	4	–65	795	792	320	535	1120	21	11	3.50	2.09	1.91	0.10
2. Sugarloaf Mountain	4	170	760	762	440	915	1980	25	5	4.50	2.16	5.00	0.40
3. Sugarloaf Mountain	4	165	825	713	470	915	1600	30	7	3.40	1.75	4.29	0.34
4. Sugarloaf Mountain	4	75	795	689	440	1065	1675	31	6	3.81	1.57	5.17	0.39
5. Crocker Mountain	4	75	885	853	425	760	1450	27	7	3.41	1.91	3.86	0.23
6. Crocker Mountain	3	80	915	847	275	760	1220	22	7	4.44	1.61	3.14	0.13
7. Crocker Mountain	4	–85	900	817	425	1370	2590	27	6	6.09	1.89	4.50	0.75
8. Crocker Mountain	4	–10	825	725	395	760	1370	22	8	3.47	1.80	2.75	0.21
Black Nubble	4	–125	865	860	290	990	2135	19	4	7.36	2.16	4.75	0.31
Tim Mountain	4	–120	795	750	350	1850	2745	18	3	7.84	1.48	6.00	0.89
average	**3.9**	**16.0**	**836.0**	**780.8**	**383.0**	**992.0**	**1788.5**	**24.2**	**6.4**	**4.78**	**1.84**	**4.14**	**0.37**
standard deviation	**0.3**	**111.9**	**52.3**	**62.1**	**69.2**	**375.1**	**560.0**	**4.5**	**2.2**	**1.70**	**0.24**	**1.24**	**0.26**
												total volume	**3.75**
Green Moutains, Vt.				**RBW/PTD**									
Miller Brook valley	3	135	395	360	715	1675	3505	25	2	4.90	2.09	12.50	2.10
Ritterbush valley	3	115	360	335	220	1830	2135	18	2	9.70	1.17	9.00	0.43
Belvidere valley	3	–105	360	365	190	1525	2440	11	1	12.84	1.60	11.00	0.35
1. Mansfield	3	–125	825	1005	610	950	2200	25	12	3.61	2.32	2.08	0.64
2. Mansfield	4	–105	915	945	640	1900	2400	29	11	3.75	1.26	2.64	1.46
3. Mansfield	3	–50	795	795	760	1100	2300	27	13	3.03	2.09	2.08	0.96
4. Mansfield	3	–55	795	795	605	1200	2150	24	12	3.55	1.79	2.00	0.78
5. Mansfield	3	110	975	975	700	1700	2450	37	13	3.50	1.44	2.85	1.46
Mansfield 1 - 5 average	**3.2**	**–45.0**	**861.0**	**903.0**	**663.0**	**1370.0**	**2300.0**	**28.4**	**12.2**	**3.49**	**1.78**	**2.33**	**1.06**
standard deviation	**0.4**	**92.4**	**80.5**	**100.8**	**66.1**	**408.7**	**127.5**	**5.2**	**0.8**	**0.27**	**0.44**	**0.39**	**0.38**
												total volume	**5.30**
Adirondack Mountains, N.Y.													
White Brook valley	3	65	900		325	1250	3050	22	9	9.38	2.44	2.44	0.62
Catskill Mountains, N.Y.													
Johnson Hollow Brook valley	2	50	700		370	1220	1980	24	7	5.35	1.62	3.43	0.45

[1] Grade follows classfication of Evans and Cox (1995), whereby 1 = classic, with all textbook attributes, 2 = well-defined, with headwall and floor clearly developed and headwall curves around cirque floor, 3 = definite, with no debate over cirque status, but one characteristic may be weak, 4 = poor, some doubt, but well-developed characteristics compensate for weak ones, 5 = marginal, with cirque status and origin doubtful.

[2] Aspect is direction faced by central headwall perpendicular to long axis of cirque measured to nearest 5° azimuth (negative values increase from 360° to 180°).

[3] Schrund altitudes in left column measured as most obvious break in slope denoted by contour lines, following method of Evans and Cox (1995), whereby headwall slopes generally more than 35° and floor slopes generally less than 20°, with limit between the two at 27°; schrund altitudes in right column measured by method of Goldthwait (1970), with initials of workers for particular areas as explained in text.

[4] Height measured from average top of headwall to lowest floor altitude (to nearest 5 m).

[5] Average width determined by numerous measurements from top of sidewall to top of opposite sidewall along long axis of cirque (to nearest 5 m).

[6] Length measured from top of headwall to cirque mouth, or where sidewalls abruptly end or drop in altitude (to nearest 5 m).

[7] Average headwall slope measured from top of headwall to cirque mouth, or where sidewalls abruptly end or drop in altitude (to nearest 5 m).

[8] Average floor inclination measured below schrund altitude.

[9] Volume calculated by: (height × width × length) / 2.

L:H ratio of 2.70:1 for Presidential Range cirques (n = 17) compares with averages of 2.8:1 to 3.2:1 for over 400 cirques in English Lake District (Embleton and King, 1975) and a median of 4.29:1 for 165 cirques on Baffin Island (Andrews and Dugdale, 1971). The mean L:W ratio of 1.64:1 for Presidential Range cirques compares with a median of 1.3:1 for the same 165 cirques on Baffin Island. The average volume of material removed from Presidential Range cirques is 0.25 km^3, with a total 4.25 km^3. Using a range of empirically-derived cirque erosion rates (Anderson, 1978), about 5.9×10^4 to 10.4×10^6 yrs would be required to erode the average Presidential Range cirque, assuming that 100% of the cirque forms are created by glacial erosion.

Average headwall slopes summarized in this report (mean 33.6°, Table I) are naturally lower than the *steepest* headwall slope segments (mean 37°) measured for the Presidential Range by Goldthwait (1970, Table 1, p. 89), although the average inclinations of floors below schrund altitudes are similar (11.2° vs 10°, respectively). Only one cirque floor ("Sphinx Basin") has an inclination (21°) higher than the upper limit (20°) suggested by Evans and Cox (1995); however, the majority of Presidential Range cirques have average headwall slopes a few degrees less than their suggested lower limit (35°). All Presidential Range cirques have headwall-to-floor slope ratios below 4.7:1, except for Upper Oakes Gulf (10.67:1; average floor inclination 3°); the standard deviation of 2.08 drops in half if the Upper Oaks Gulf ratio is not included. Thus, using new maps and different methods for measuring schrund altitudes and headwall slopes do not result in headwall-to-floor slope ratios much different than those determined by Goldthwait (1970) for Presidential Range cirques.

MOUNT MOOSILAUKE, NEW HAMPSHIRE

Morphometric data for four cirques on Mount Moosilauke (Figs. 1, 4) in the western White Mountains of New Hampshire are summarized in Table I. In comparison with cirques in the Presidential Range, cirques on Mount Moosilauke average the same grade (2.8), have a similar high variability in aspect (two with azimuths southeast; two with azimuths north-northeast), average almost 250 m lower in schrund altitudes (although average about 165 m lower if only highest three cirques on Mount Moosilauke considered), have higher length-to-height and length-to-width ratios (4.40:1 vs. 2.70:1 and 2.47:1 vs. 1.64:1, respectively), have lower average headwall and floor slope angles but similar average headwall-to-floor slope ratios (3.24 vs. 3.62), and have almost identical average volumes (0.24 vs. 0.25 km^3 per cirque). Measured schrund altitudes for the three highest cirques on Mount Moosilauke are about 85 m lower (1065 m vs. 1150 m) than measured by Haselton (1975).

MOUNT KATAHDIN, MAINE

Morphometric data for eight cirques on Mount Katahdin (Figs. 1, 5) in west-central Maine are summarized in Table I. In comparison with cirques in the Presidential Range, those on Mount Katahdin average higher in grade (1.6 vs. 2.8, with four ranked grade 1), have a similar high variability in aspect (four with azimuths west or northwest; four with azimuths east or northeast), average over 240 m lower in schrund altitudes, have higher average length-to-height and length-to-width ratios (3.98:1 vs. 2.70:1 and 2.12:1 vs. 1.64:1, respectively), have very similar average headwall slopes (34.3° vs. 33.6°) but a higher average headwall-to-floor slope ratio (6.25:1 vs. 3.62:1), and have similar average volumes (0.28 vs. 0.25 km^3 per cirque, albeit a 0.21 km^3 standard deviation for each group), with a total cirque volume about half of the Presidential total, reflecting about half the number of cirques.

The slightly different schrund altitudes for Katahdin's cirques in the right column labeled "PTD" in Table I were measured from a 15-minute topographic map (Davis, 1976), as opposed to measurements in the left column that were made on a 7.5-minute map; however, the contour interval of 20 ft (6 m) was the same on both maps. One might be tempted to explain the lower schrund altitudes on Mount Katahdin as indication of greater snow-line depression than in the Presidential Range; however, different underlying bedrock (Devonian granite on Mount Katahdin [Osberg *et al.*, 1985] vs. Silurian-Devonian schist, gneiss, and quartzite of the Rangeley, Perry Mountain, Small Falls, Madrid, and Littleton Formations in the Presidential Range [Eusden *et al.*, 1996]) might also be an important controlling factor for schrund altitude (*e.g.*, Evans, 1994). In Mount Katahdin's three large east-facing cirques, the steep headwalls appear to be controlled by the prominent vertical jointing in the granite bedrock and the consequent release of large exfoliation sheets.

LONGFELLOW MOUNTAINS, MAINE

Morphometric data for 10 cirques in the Longfellow Mountains in west-central Maine (Figs. 1, 6) are summarized in Table I. In comparison with cirques of the Presidential Range, Mount Moosilauke, and Mount Katahdin, cirques in the Longfellow Mountains average lower in grade, with nine of 10 cirques grading 4. Cirque aspect in the Longfellow Mountains is more variable and schrund altitudes average about 390, 150, and 160 m lower than in the Presidential Range, on Mount Moosilauke, and on Mount Katahdin, respectively. Schrund altitudes measured from 7-5 minute topographic maps average about 55 m higher than schrund altitudes measured by Borns and Calkin (1977) from 15-minute maps (labeled "HWB/PEC" in Table I), although the contour intervals were the same at 20 ft (6 m).

Although average height is smaller, average width and length are greater, hence average cirque volume is greater for Longfellow than for Presidential, Moosilauke, or Katahdin cirques. Average headwall slopes are less steep than those for cirques of the Presidential Range and Mount Katahdin, but similar to those on Mount Moosilauke. Concomitant lower floor inclinations result in an average headwall-to-floor slope ratio between that for cirques of the Presidential Range, Mount Moosilauke, and Mount Katahdin (4.14:1 vs. 3.62:1, 3.24:1, and 6.25:1, respectively). There does not appear to be a relationship between cirque morphometry and bedrock in the Longfellow Mountains (Table I), as cirques with similarly steep headwalls are cut into a variety of bedrock types (Osberg *et al.*, 1985), including Devonian granodiorite (Crocker 6 and 7 and Black Nubble cirques), Devonian gabbro and ultramafics (Crocker 5 and 8 and Sugarloaf 1 cirques), and pelites of the Silurian Perry Mountain and Rangeley Formations (Sugarloaf 2, 3, and 4, and Tim Mountain cirques).

GREEN MOUNTAINS, VERMONT

Waitt and Davis (1988, Table 2. p. 508) summarized data for two groups of cirques in the Green Mountains (Fig. 1), one group consisting of three low-altitude basins (Figs. 7, 8), the other seven high-altitude cirques in the Mount Mansfield area (Fig. 9). Of the seven high-altitude cirques, Waitt and Davis (1988) considered two to be questionable in origin, so data are only presented here for the five cirques on Mount Mansfield (Fig. 9, Table I). Because the three low-altitude basins led Wagner (1970, 1971) and Connally (1971) to suggest that local glaciers persisted in the Green Mountains following dissipation of continental ice, these are considered first. These three basins are unusual in that two hold very large lakes compared with the typical size of tarns, two are oriented to the southwest, they average about 370 m in schrund altitudes, one has a volume almost three times larger than any other cirque in the northeastern United States, and headwalls are about half the slope angle of cirques in the Presidential Range and Mount Katahdin. These characteristics, along with odd digitate shapes (Figs. 7, 8), led Waitt and Davis (1988) to question whether the basins were cirques at all, but rather to interpret them as valley heads last occupied by late-glacial tongues of continental ice.

Morphometric data for the five high-altitude cirques on Mount Mansfield in the northern Green Mountains (Figs. 1, 9) are summarized in Table I. In comparison with cirques of the Presidential Range, Mount Moosilauke, Mount Katahdin, and the Longfellow Mountains, the high-altitude cirques on Mount Mansfield are similar in grade, with an average of 3.2. Four of the high-altitude Mansfield cirques have a westerly aspect, with an average of northwest for the five cirques. Schrund altitudes average about 55 m higher than in the Longfellow Mountains, but about 270, 145, and 150 m lower than cirques in the Presidential Range and Mount Moosilauke, and Mount Katahdin, respectively. With the exception of Mansfield cirque #1, earlier measurements of schrund altitudes reported in Waitt and Davis (1988; labeled “RBW/PTD” in Table I) are similar to the new measurements in Table I. Most other morphometric measurements for the high-altitude cirques on Mount Mansfield are similar to those for cirques of the Presidential Range, Mount Moosilauke, Mount Katahdin, and the Longfellow Mountains; however, the Mansfield cirques are 3 to 4 times larger in average volume. In summary, the high-altitude cirques on Mount Mansfield appear to have morphometric characteristics that are similar to those of other cirques in northern New England, but have less in common with the three low-altitude cirques or valley heads in the northern Green Mountains.

ADIRONDACK MOUNTAINS, NEW YORK

Craft (1976) identified 224 cirque basins in the Adirondacks and proposed a number of sites as locations for cirque glaciers persisting after wastage of continental ice. Warburton (1982) noted that many cirques in the Adirondacks have breached headwalls, whereas other basins have been so severely eroded by valley glaciers that they have lost much of their cirque form, a likely indication that erosive continental ice post-dated cirque glaciation in the High Peaks region. Based on field mapping, Barclay (1993) rejected a number of Craft’s (1976, 1979) sites as potential locations of cirque glaciers post-dating continental ice recession; however, Donald Rodbell (oral commun., 1998) suggests that many cirques in the Adirondacks have moraines, including some with tarns that could yield sediment cores and minimum-limiting radiocarbon ages for deglaciation.

Morphometric data for only one Adirondack site, White Brook valley on the northeast side of Whiteface Mountain (Figs. 1, 10), is presented in Table I. This grade 3 cirque has a northeast aspect, a schrund altitude between the average schrund altitudes for cirques in the Longfellow Mountains and on Mount Katahdin, and a greater length than the average cirque in the Presidential Range, on Mount Katahdin, and in the Longfellow Mountains. However, the average headwall slope is only 22°, similar to the low average headwall slope angles in the Longfellow and Green Mountains.

CATSKILL MOUNTAINS, NEW YORK

Johnson Hollow Brook valley is one of many cirques in the Catskills proposed to be sites for local glaciers that flowed independently after recession of continental ice (Rich, 1906, 1935; Johnson, 1917). Because Johnson Hollow Brook valley (Figs. 1, 11) is oriented toward the northeast, it is ideally suited to maximize wind drifted snow, to minimize solar radiation, and to provide ice-flow indicators opposite those of continental ice. The cirque is similar in size to the larger cirques in the White Mountains, on Mount Katahdin, and in the Longfellow Mountains; however, its average headwall slope (24°) is about 10° lower and its schrund altitude (700 m) is between 300 and 500 m lower than the average cirque on Mount Katahdin and in the Presidential Range, respectively.

DISCUSSION

Using an atmospheric lapse rate (6°C/1km; *e.g.*, Barry, 1992, p 45) and assuming a current July freezing isotherm at 3050 m altitude in northern New England, Goldthwait (1970) and Davis (1976) estimated that cirque glaciers in the Presidential Range and on Mount Katahdin would have required summer temperature depressions about 9°C and 13°C below modern-day values, respectively. Using a 5.3°C/1km lapse rate, Loso *et al.* (1998) calculated a summer temperature depression about 13.6°C below modern-day values to support a local glacier in the Miller Brook valley of the Green Mountains, assuming an equilibrium-line altitude of 480 m (using the 6°C/1km lapse rate and the 395 m schrund altitude measured in Table I would yield a summer temperature depression about 16°C). Using an atmospheric lapse rate of 6°C/1km and assuming a current July freezing isotherm at 3350 m altitude in southern New York would require a summer temperature depression about 16°C below modern-day values to support a cirque glacier at the head of Johnson Hollow Brook valley in the Catskills. However, these estimates do not consider increased winter precipitation (Leonard, 1989), which could compensate for part of the required temperature depression. For example, Lederer (1998) suggested that winter precipitation between 12,000 and 11,000 yrs BP may have increased in the Catskills because of the proximity of glacial Lake Iroquois (Pair and Rodrigues, 1993). Moreover, favorable cirque aspect for shielding from summer insolation and enhancement of wind-drifted snow from prevailing winds (Goldthwait, 1970) also may have influenced snow accumulation in cirques. Indeed, Havens (1960) noted that large snow banks commonly remained throughout the summer on the floor of Upper Tuckerman Ravine in the Presidential Range during the early 1900s. However, lichenometric data, similar to that collected on the summit of Mount Washington by Mayewski and Jeschke (1978), suggest that snow banks in Tuckerman Ravine did not persist or increase in size sufficiently to form a cirque glacier during Neoglaciation (the past 2000 to 3000 years).

If cirque glaciers did persist after recession of continental ice in the northeastern United States, the most likely time for their occurrence would have been during the Younger Dryas (YD) cooling event (Alley *et al.*, 1993; Mayewski *et al.*, 1993), which is radiocarbon-dated about 11,000 to 10,000 yrs BP. The earlier Killarney Oscillation, a cooling event radiocarbon-dated between 11,300 and 11,000 yrs BP (Levesque *et al.*, 1994), was probably too short-lived to support cirque glaciers and much of the northeastern United States remained under continental ice during the Heinrich I cooling event (Bond *et al.*, 1992) about 14,000 yrs BP. Moraines that may have been formed by cirque glaciers are not present in the Presidential Range, Longfellow Mountains, or Adirondacks. Moraines that flank the east and south slopes of Mount Katahdin were formed by continental ice surrounding the mountain, and not by cirque glaciers (Davis, 1976, 1978, 1983, 1989). Radiocarbon ages from pond and bog-bottoms on Mount Katahdin are far too young to provide closely-limiting ages for these moraines (Davis and Davis, 1980); however, one preliminary set of cosmogenic ^{10}Be and ^{26}Al exposure ages for a boulder on one of the moraines suggests an age older than the YD (Davis and Bierman, unpublished data). Cosmogenic radionuclides have provided late-glacial moraine chronologies in many areas of the world (*e.g.*, Gosse *et al.*, 1995a, b). Although most of these areas occur at higher altitudes where cosmogenic radionuclide production rates are higher than in the northeastern United States, the sensitivity of the method is improving and can already resolve late-glacial chronologies at sea level (*e.g.*, Davis *et al.*, 1999).

A moraine-like ridge in the Miller Brook valley in the Green Mountains (Wagner, 1970, 1971; Connally, 1971) is now believed to be the result of late-glacial tongues of continental ice (Waitt and Davis, 1988; Loso *et al.*, 1998) or to be an esker (Wright et al., 1997a, b). Moreover, the cross-valley moraines in Ritterbush valley thought to be as young as 10,800 yrs BP (Sperling *et al.*, 1989) are now known to be more than 12,000 year old, or pre-YD in age, based on radiocarbon ages of sediments from Ritterbush Pond (Bierman *et al.*, 1997; Lin *et al.*, 1995; Lini *et al.*, 1995).

At Johnson Hollow Brook valley in the Catskills, a moraine at the mouth of Schoendorf cirque, long held to be deposited by a local glacier persisting after dissipation of continental ice (Rich, 1906, 1935; Johnson, 1917), was recently shown to be as young as 10,860 yrs BP (Lederer, 1998; Lederer and Rodbell, 1998). Thus, the Catskills stand alone as a location in the northeastern United States where cirque glaciers post-date recession of continental ice, and possibly formed during the early part of the YD. The nearest areas where glacial geological data provide similar evidence for local glaciation post-dating wastage of continental ice are the Maritime Provinces of Canada, where lowland ice expanded during the YD (Stea and Mott, 1989). In northern Maine, lithological analyses and radiocarbon-dating of lake sediment cores also suggest that a local lowland icecap may have persisted into YD time (Dorion, 1998). If local ice were present in Catskill cirques following dissipation of continental ice, one must ask why we do not see evidence for local glaciers surviving continental ice recession in higher-altitude and higher latitude cirques in Maine and New Hampshire.

In the Maritime Provinces, the climatic signal for the YD is pronounced in a variety of lake sediment proxy records, including loss on ignition, sediment grain size, pollen, plant macrofossils, and chironomids (Mayle *et al.*, 1993a, b; Levesque *et al.*, 1994, 1997). In the northeastern United States, the climatic signal for the YD is less pronounced but is still recognizable in LOI, pollen, and macrofossil records from lake sediment cores in northwestern Maine and northern New Hampshire (Thompson *et al.*, 1996, this volume; Dorion, 1998). Elsewhere in the northeastern United States, the YD signal is more subtle in lake sediments, but is nevertheless present in most pollen records (Peteet *et al.*, 1990). For example, in pollen records from lake sediment cores on the western flank of the Adirondacks (Davis *et al.*, 1995a), in southern Vermont (Davis *et al.*, 1995b), in northern Vermont (Lin *et al.*, 1995; P.T. Davis *et al.*, unpublished data), and in northwestern New Hampshire (Likens and Davis, 1975; Davis *et al.*, 1980; Davis and Ford, 1982), a decrease is seen in oak pollen with a concomitant increase in alder pollen during an interval that is radiocarbon dated between about 11,000 and 10,000 yrs BP. To interpret such paleoclimatic signals, one must sample at a resolution high enough to recognize short-lived events, which is now underway for plant macrofossils, pollen, chironomids, and other proxies in many high-altitude ponds in the White Mountains (Ray Spear and Les Cwynar, oral commun., 1998). Unfortunately, few cirques in the northeastern United States are blessed with tarns suitable for such paleoclimatic studies.

CONCLUSIONS

Morphometric analysis suggests that cirques in the Presidential Range in New Hampshire and Mount Katahdin in Maine should be the most likely settings for survival or rejuvenation of local glaciers after icesheet recession. However, evidence for cirque glaciers persisting after wastage of continental ice remains lacking for the Presidential Range and Mount Moosilauke of New Hampshire, Mount Katahdin and the Longfellow Mountains of Maine, and the Adirondack Mountains of New York. Radiocarbon dating of lake sediment cores in the northern Green Mountains of Vermont and in the Catskill Mountains of New York has produced additional chronological data that address the presence of cirque glaciers following dissipation of continental ice in the northeastern United States.

At Ritterbush Pond in northern Vermont, radiocarbon ages from sediment cores suggest that a low-altitude valley head located up-valley of a series of cross-valley moraines was ice-free by 11,940 ^{14}C yrs BP, thus precluding a glacier advance during the Younger Dryas cold event. In Miller Brook valley in northern Vermont, a cross-valley ridge that was thought to be a moraine formed by a cirque glacier is now believed to be an esker. Although some workers argue that these ridges in Vermont are evidence for cirque glaciation, morphometric analysis that includes an average schrund altitude of about 370 meters for these low-elevation valley heads suggests that these landforms were more likely formed by a tongue of continental ice during Late Wisconsinan deglaciation.

At Johnson Hollow Brook valley in the Catskills of New York, a radiocarbon age from basal sediments in a tarn dammed by a moraine suggests that the cirque was not ice-free until 10,860 ^{14}C yrs BP, thus allowing the possibility of a local glacier persisting during the earliest part of the Younger Dryas cold interval. However, cirque morphometric data suggest that a schrund altitude of 700 meters for the cirque at the head of Johnson Hollow Brook valley is much less conducive for persistence of a local glacier than are cirques with similar aspect and much higher schrund altitudes in the Presidential Range and on Mount Moosilauke in New Hampshire and on Mount Katahdin in Maine. This dichotomy can only be resolved by further coring and radiocarbon dating of sediments in ponds and bogs in alpine areas of the northeastern United States.

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APPENDIX 1

TOPOGRAPHIC MAPS USED FOR CIRQUE MORPHOMETRIC ANALYSES

Presidential Range, New Hampshire

All cirque morphometric measurements were made from Bradford Washburn's *Mount Washington and the heart of the Presidential Range, New Hampshire*, 1:20,000-scale topographic map, with a contour interval of 50 feet and intermediate contours of 25 feet, surveyed and edited between 1978 and 1987, produced by Boston's Museum of Science in 1988, published and distributed by the Appalachian Mountain Club, Boston, Mass. 02108. With thousands of laser theodolite survey measurements, this map is far more accurate than any other topographic maps for the Presidential Range. Contour lines are easier to read on later editions, in comparison with the first edition, which used a dark green background pattern. Morphometric measurements of Presidential Range cirques #1-16 (Table I) were also made from the U.S. Geological Survey's *Mt Washington, New Hampshire* 7.5- × 15-minute, 1:25,000-scale metric topographic map, with a contour interval of 6 meters. This map was compiled by photogrammetric methods from aerial photographs taken in 1972, field checked in 1975, and edited in 1982. Morphometric measurements of Presidential Range cirques #14-17 were also made from the U.S. Geological Survey's *Stairs Mtn. Quadrangle, New Hampshire*, provisional edition 7.5-minute, 1:24,000-scale topographic map, with a contour interval of 40 feet. This map was compiled from aerial photographs taken in 1981, field checked in 1983, and edited in 1987.

Mount Moosilauke, New Hampshire

All cirque morphometric measurements were made from the U.S. Geological Survey's *Mt. Moosilauke Quadrangle, New Hampshire*, 7.5-minute, 1:24,000-scale topographic map, with a contour interval of 40 feet. This map was compiled from aerial photographs taken in 1964 and field checked and edited in 1967.

Mount Katahdin, Maine

All cirque morphometric measurements were made from the U.S. Geological Survey's *Mount Katahdin Quadrangle, Maine*, provisional edition 7.5-minute, 1:24,000-scale topographic map, with a contour interval of 20 feet. This map was compiled from aerial photographs taken in 1982, field checked in 1985, and edited in 1988.

Longfellow Mountains, Maine

Morphometric measurements for cirques #1-6 (Table I) were made from the U.S. Geological Survey's *Sugarloaf Mtn., Maine, Quadrangle*, provisional edition 7.5-minute, 1:24,000-scale topographic map, with a contour interval of 20 feet. Measurements for cirques #5-8 and Black Nubble cirque (Table I) were made from the U.S. Geological Survey's *Black Nubble, Maine, Quadrangle*, provisional edition 7.5-minute, 1:24,000-scale topographic map, with a contour interval of 20 feet. These maps were compiled from aerial photographs taken in 1985, field checked in 1987, and edited in 1989. Measurements for Tim Mountain cirque (Table I) were made from the U.S. Geological Survey's *Tim Mountain, Quill Hill, Black Mountain, and Kennebago Lake, Maine Quadrangles*, 7.5-minute, 1:24,000-scale topographic maps, with a contour interval of 20 feet. These maps were compiled from aerial photographs taken in 1966, and field checked and edited in 1969 and 1970.

Green Mountains, Vermont

Morphometric measurements for Miller Brook valley cirque were made from the U.S. Geological Survey's *Bolton Mountain, Vt., Quadrangle*, 7.5-minute, 1:24,000-scale topographic map, with a contour interval of 20 feet. This map was compiled from aerial photographs taken in 1947, field checked and edited in 1948, and photo-inspected in 1983. Morphometric measurements for Belvidere and Ritterbush valley cirques were made from the U.S. Geological Survey's *Eden and Hazens Notch, Vermont, Quadrangles*, provisional edition 7.5-minute, 1:24,000-scale topographic maps, with a contour interval of 6 meters. These maps were compiled from aerial photographs taken in 1980 and 1981, field checked in 1982, and edited in 1986. Morphometric measurements for the Mount Mansfield cirques were made from the U.S. Geological Survey's *Mount Mansfield, Vermont, Quadrangle*, 7.5-minute, 1:24,000-scale topographic map, with a contour interval of 20 feet. This map was compiled from aerial photographs taken in 1947 and 1978, field checked in 1948, photorevised and edited in 1980, and photoinspected in 1983.

Adirondack Mountains, New York

Morphometric measurements for White Brook valley cirque were made from the U.S. Geological Survey's *Wilmington, New York*, 7.5- × 15-minute, 1:25,000-scale metric topographic map, with a contour interval of 10 meters. This map was compiled from aerial photographs taken in 1976, field checked in 1976, and edited in 1978.

Catskill Mountains, New York

Morphometric measurements for Johnson Hollow cirque were made from the U.S. Geological Survey's *Prattsville and Roxbury, New York, Quadrangles*, 7.5-minute, 1:24,000-scale, with a contour interval of 20 feet. These maps were compiled from aerial photographs taken in 1943, and field checked and edited in 1945; the Roxbury quadrangle was also photo-inspected in 1981.