Deglaciation of the Upper Androscoggin River Valley and Northeastern White Mountains, Maine and New Hampshire

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ABSTRACT

The mode of deglaciation of the White Mountains of northern New Hampshire and adjacent Maine has been a controversial topic since the late 1800's. Recent workers have generally favored regional stagnation and downwastage as the principal means by which the late Wisconsinan ice sheet disappeared from this area. However, the results of the present investigation show that active ice persisted in the upper Androscoggin River valley during late-glacial time. An ice stream flowed eastward along the narrow part of the Androscoggin Valley between the Carter and Mahoosuc Ranges, and deposited a cluster of end-moraine ridges in the vicinity of the Maine-New Hampshire border. We have named these deposits the "Androscoggin Moraine." This moraine system includes several ridges originally described by G. H. Stone in 1880, as well as other moraine segments discovered during our field work. The ridges are bouldery, sharp-crested, and up to 30 m high. They are composed of glacial diamictons, including flowtills, with interbedded silt, sand, and gravel. Stone counts show that most of the rock debris comprising the Androscoggin Moraine was derived locally, although differences in provenance may exist between moraine segments on opposite sides of the valley.

Meltwater channels and deposits of ice-contact stratified drift indicate that the margin of the last ice sheet receded northward. Early penetration of mountain passes by esker-forming glacial streams was accompanied and followed by cutting of meltwater channels and deposition of ice-contact fluvial and lacustrine sediments as the uplands were uncovered. Lakes were ponded between the ice margin and north- to west-facing slopes. During the late stage of deglaciation, larger volumes of meltwater deposits accumulated on valley floors.

Most of the features formed by glacial meltwater in the study area could have developed in association with stagnant ice. However, the Androscoggin Moraine, together with moraines described by R. F. Gerath in the Berlin-Gorham area, supports the continued presence of active ice in the upper Androscoggin River basin when the Presidential, Carter, and Mahoosuc mountain ranges had emerged from the late Wisconsinan ice sheet. Glacial striation trends and previous studies in the White Mountains suggest that an ice lobe spilled into the Androscoggin River valley from the Connecticut River basin to the north and west. On the basis of the limited available radiocarbon data, the Androscoggin Moraine is believed to have been deposited by the Laurentide ice sheet at about 14,000 yr B.P.
INTRODUCTION

The manner in which the White Mountains were deglaciated has long been one of the most debated issues in the study of New England glacial geology. Much of the research in this region, especially in New Hampshire, has been directed toward the following topics: the configuration of the receding late Wisconsinan ice margin, the scale and relative importance of stagnation versus active ice flow in the marginal zone of the waning ice sheet, and whether an ice cap or alpine glaciers persisted in the mountains subsequent to their emergence from the continental ice. The theories concerning these problems have been summarized by Goldthwait (1916), Goldthwait and Mickelson (1982), Gerath et al. (1985), and Waite and Davis (1988). Gerath et al. (1985) pointed out that much detailed geologic mapping remains to be done to improve our understanding of White Mountain glacial history. In the northern part of the mountains, the only intensive mapping studies are those by Goldthwait (1940, 1970) in the Presidential Range and Gerath (1978) in the Berlin-Gorham area.

This report discusses the style of deglaciation in the northeastern White Mountains and adjacent western Maine. A review and synthesis of previous work are followed by the results of our investigations in a section of the Androscoggin River valley that straddles the Maine-New Hampshire border. The principal study area extends from Bethel, Maine, west and north to the origin of the Androscoggin River at the outlet of Umbagog Lake (Fig. 1).

Recent surficial geologic mapping by W. B. Thompson in the Maine portion of the Androscoggin River valley led to the discovery of a large cluster of glacial end moraines in the vicinity of the state line. Part of this moraine system had been briefly described by Stone (1880, 1899) during his work for the U.S. Geological Survey. Thompson (1983, 1986) concurred with Stone’s identification of the ridges as moraines and assigned the name "Androscoggin Moraine" to this group of deposits.

The Androscoggin Moraine is significant for two reasons. First, it provides information about the style of deglaciation in the mountains along the Maine-New Hampshire border. The moraine may indicate persistence of active ice in the upper Androscoggin and Connecticut River basins when flow had ceased elsewhere in the mountains. Second, the Androscoggin Moraine appears to be the most clearly defined of the few moraines that have been described in the White Mountains and is the only known example of a cross-valley moraine. Problems discussed here include the age of the moraine system and whether it was deposited by the Laurentide ice sheet or by a northern Appalachian ice mass subsequent to marine incursion in the St. Lawrence Lowland. We also consider ice-flow directions indicated by glacial striations and streamlined hills, end moraines reported by previous workers, the sequence of meltwater deposits, and the relationships of these features to the regional history of deglaciation in the mountains of northern New Hampshire and western Maine.

PREVIOUS STUDIES IN THE UPPER ANDROSCOGGIN RIVER BASIN AND ADJACENT AREAS

Until recent years, little detailed work was done on the glacial geology of the upper Androscoggin River basin. Hitchcock (1878b) mapped glacial striations and "modified drift" (water-laid deposits) in New Hampshire, producing a series of rudimentary surficial geologic maps encompassing the whole state. During the late 1800’s and early 1900’s, most research centered on the possible existence of an ice cap or system of alpine valley glaciers in the higher parts of the White Mountains.

Packard (1867) envisioned an ice cap that "discharged local glaciers into the principal valleys radiating from the central peaks." He thought that one of these glaciers flowed northward down the Peabody River valley (Fig. 1) and joined the "Androscoggin glacier" at Gorham. From observations on striation trends, Vose (1868) expressed the same opinion concerning the ice flow direction in the Peabody Valley. Hitchcock (1878a) refuted this theory on the basis of stoss-and-lee erosional forms on bedrock, which indicated ice flowing southward up the Peabody Valley. Vose also tested the possibility of local glaciation in the Androscoggin River valley by measuring striation trends in the east-west section of the valley between Bethel and Gorham (Fig. 1). He found striations indicating eastward flow and concluded that a valley glacier had extended at least to Bethel. Vose distinguished the proposed valley glaciation from the "general operation" (presumably of continental ice) upon all of New England, but acknowledged that the relationship between these events was unknown.

Advocates of local glaciation in the White Mountains subsequent to emergence from the most recent continental ice sheet included Agassiz (1870a,b), Hitchcock (1878a), and Upham (1904). These geologists described the Bethlehem Moraine, and inferred that it was deposited by ice flowing from the Franconia Range northward into the Ammonoosuc River valley. This complex cluster of moraine segments is located east of Littleton, New Hampshire, about 20-30 km west of the present study area. Upham (1904) claimed that the Bethlehem Moraine was formed adjacent to an ice cap, rather than the valley glacier suggested by Agassiz and Hitchcock. Goldthwait (1916) pointed out numerous flaws in the local-ice theories, and the vague, inconclusive nature of the evidence that had been advanced in support of this model. However, the possibility of a late-glacial residual ice cap on the higher peaks of the White Mountains was entertained as recently as the 1950’s (Flint, 1951).

The Bethlehem Moraine was also at the center of early debate about whether deglaciation occurred by regional stagnation or the northward retreat of an active ice margin. Goldthwait (1938) concluded that widespread stagnation and downwasting was the chief mode of glacial retreat from the White Mountains. He relegated the Bethlehem Moraine to being just "a zone of massive kettled outwash." On the other hand, Lougee (1940) emphatically upheld the Bethlehem deposits as moraine ridges,
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Figure 1. Map showing locations of principal geographic features and end moraines mentioned in text. AM: Androscoggin Moraine system (outer boundary). CM: Copperville Moraine. CPM: C Pond Moraine. SM: Success Moraine.
his chief evidence apparently being the morphology and bouldery surfaces of these deposits. Lougee agreed with Crosby (1934) that the moraine system was produced by a glacial readvance. The readvance theory was supported by exposures in the Connecticut River valley west of Littleton showing two tills separated by thrust-faulted sand and gravel.

The concept that widespread ice stagnation and downwastage began early in the deglaciation of the White Mountains and was the chief mode of ice retreat has generally prevailed in recent years. Goldthwait and Mickelson (1982) advocated this model on the strength of similarities between glacial features in the White Mountains and those of the Glacier Bay area in southeast Alaska. Much of the recent research in the White Mountains has centered on determining the time of development of Presidential Range cirque glaciers relative to overriding of the mountains by the late Wisconsinan continental ice sheet (Goldthwait, 1940, 1970; Bradley, 1981, 1982; Gerath and Fowler, 1982; Fowler, 1984; Waitt and Davis, 1988). The cirque controversy is not central to our discussion of the upper Androscoggin River valley region and thus will not be considered further.

On the Maine side of the border, north and northeast of the Androscoggin Moraine, Leavitt and Perkins (1935) briefly described glacial features in the Mahoosuc Range. They proposed that a high-level ice-contact lake drained eastward through Mahoosuc Notch (south of Mahoosuc Mountain, Fig. 1) as the glacier margin began to withdraw from these mountains. Leavitt and Perkins also named glacial Lake Cambridge, which they described as first draining south through the Grafton Notch spillway. This lake had several lower outlets as the ice-margin receded and finally emptied westward into the upper end of the Androscoggin River valley.

Caldwell (1974) reported on the results of reconnaissance surficial mapping in western Maine. He noted the presence of an end moraine (which he named the "C Pond Moraine") in a narrow pass through the northeastern Mahoosuc Range (Fig. 1). Caldwell described this moraine as being hummocky, strewn with large boulders, and incised by a 6 m-deep meltwater channel.

**ICE FLOW DIRECTIONS IN THE UPPER ANDROSCOGGIN BASIN**

Two types of ice flow indicators have been recorded in the study area: glacially streamlined hills, and striations and grooves on bedrock outcrops. Figure 2 shows the ice flow directions measured by earlier workers and the present authors. The striations and grooves usually indicate ice flow trends, rather than absolute directions. However, roche moutonnée and flag-and-tail forms demonstrate that the regional flow of glacial ice was toward the southeast or east. In most places this information enabled a reasonably certain flow direction to be assigned to the striations.

The streamlined hills are composed of till and bedrock in varying proportions. They show a wide range of sizes and include southeast-trending drumlinoid hills, crag-and-tails, and fluted till ridges. Gerath (1978) mapped large numbers of these features on the northwest side of the Mahoosuc Range in New Hampshire (Fig. 2). They also occur in adjacent western Maine, where drumlin axes trend approximately 130° (Caldwell, 1974). This belt of streamlined hills is a distinctive terrain and can be seen in high-altitude imagery (Fig. 3).

Goldthwait (1940) found that striations and grooves in the Presidential Range (southwest corner of Fig. 2) indicate an average ice flow direction of 140°. This is in accord with the general southeast trend of striations and streamlined hills throughout the present study area. However, there are significant exceptions to the regional trend, as described below.

The Androscoggin River valley abruptly narrows and curves to the east at Gorham, where it cuts across the northeast-trending mountain chain including the White Mountains and Mahoosuc Range (Fig. 1). It is evident from Figure 2 that glacial ice flow was locally channeled in conformance to this section of the valley. Numerous striations trending east to southeast have been recorded between Gorham and Bethel, commencing with the observations of Vose (1868) and Hitchcock (1878a). A prominent roche moutonnée indicating eastward ice flow can be seen adjacent to the Portland-Montreal pipeline, on the south flank of Hark Hill in Shelburne (Fig. 4). This eastward flow was topographically controlled and probably occurred as the late Wisconsinan ice sheet thinned over the mountains.

Farther up the Androscoggin River valley, between Gorham and Errol, glacial striations measured by the authors generally trend 135-155°. Thus they conform to the regional ice-flow pattern documented by Gerath (1978) and earlier workers. We had conjectured that striations in the upper Androscoggin basin might indicate a local south to southwestward flow of late-glacial ice stranded in the lowland between the mountains of northernmost New Hampshire and the Mahoosuc Range-Blue Mountains to the southeast. However, the only anomalous striations found during our reconnaissance are located on outcrops where Route 26 skirts the base of Errol Hill and Mill Mountain (Fig. 2). Here there are two outcrops that show east-trending striations in addition to the usual southeast-trending set. One site has a clearly older 103° set on the sheltered lee side of the outcrop, and a younger 135° set on the stoss surface. A similar relationship exists at the other site, where there are four striation sets that seem to indicate a progressive shift in ice-flow direction from 73° to 135°.

The eastward flow in the Errol area is thought to be unrelated to local topography, since it would have been easier for glacial ice to flow southeast along the Mollidgewock Brook valley than to cross Errol Hill and Mill Mountain. Perhaps this event was contemporaneous with the early and widespread eastward ice flow across northern Maine that was described by Lowell and Kite (1988).
Figure 2. Map showing glacial striation localities (with known or inferred ice-flow directions) and representative trends of glacially streamlined hills. Data on New Hampshire hills are from Gerath (1978); those on Maine hills are from Caldwell (1975b,c). Dots on arrows indicate sites where striations were measured; flagged arrows represent older flow directions. Sources of data: (1) Vose, 1868; (2) Hitchcock, 1878a,b; (3) Goldthwait, 1940; (4) Goldthwait et al., 1951; (5) Caldwell, 1975a,b,c; (6) Gerath, 1978; (7) measurements by the authors.
The southward deflection of ice flowing up the Peabody River valley (south of Gorham) has already been mentioned. Other topographically controlled deflections from the regional trend of glacial flow have been recorded on the north side of the Presidential Range by Goldthwait (1940). His map shows northeast-trending striations on the lower northwest slope of Mt. Adams (Fig. 2). Goldthwait also mentioned east-trending grooves on Kelton Crag, located on a northern spur of Mt. Madison. Both of these localities are on the side of Randolph Valley, which is a deep east-west trough that cuts across the divide between the Connecticut and Androscoggin River basins.

**THE ANDROSCOGGIN MORAINE**

The Androscoggin Moraine overlaps the border between Gilead, Maine, and Shelburne, New Hampshire, in the Androscoggin River valley. Altogether it consists of at least 21 moraine segments, many of which are grouped in clusters of en-echelon ridges (Fig. 4). The arcuate patterns formed by these ridges, as well as the curvature of several individual ridges, indicate that the ice margin along which they were deposited was convex down-valley. Each of the moraine segments shown in Figure 4...
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Figure 4. Surficial geologic map of the southeastern part of the Shelburne 1:24,000 quadrangle. Numbered lines indicate crests of moraine ridges (Androscoggin Moraine system).
is numbered for reference purposes in the discussion that follows.

**Previous Research**

G. H. Stone was probably the first to describe part of the Androscoggin Moraine. In an article concerning the "Androscoggin glacier" he gave a detailed morphologic description of prominent moraine ridges (9, 10, 11 on Fig. 4) on the south side of the valley (Stone, 1880). He noted that some of the morainal sediments "show signs of water-wash, with a loose structure as of gravelly upper till" (p. 300). Stone also mentioned the composite ridge (moraines 14-16) that projects southeast from Hark Hill. He illustrated this ridge in a later article, describing it as the "terminal moraine of the local Androscoggin glacier" (Stone, 1899, p. 274).

Upham (1904) briefly commented on the Androscoggin Moraine. He considered it to be possibly the most clearly defined moraine in the White Mountains, in contrast to the "promiscuous morainic belt" at Bethlehem. On the other hand, Leavitt and Perkins (1935) disputed Stone's (1899) identification of the Hark Hill ridge as an end moraine. They first described it as "a mass of bouldery till . . . deposited in the angle between the Androscoggin ice tongue and one coming down Ingalls River valley" (p. 41). Later in the same volume, Leavitt and Perkins called the deposit a kame terrace (p. 118). They claimed that no corresponding ridge exists on the south side of the Androscoggin River valley, so evidently they were unaware of Stone's 1880 article.

**Description of the Moraine System**

**Morphology.** A study of the Androscoggin Moraine was begun in 1982. The authors carried out field work in the early spring and late fall, when the absence of leaf cover increased visibility in the densely wooded terrain. The first objective was to determine whether the ridges described by Stone are in fact moraines. Some of them are oriented parallel to regional or local ice-flow directions, which suggested that they might be drift tails. Stone recognized this possibility when he investigated the ridge extending southeast from Hark Hill: "It bears N20°W, which was so near the direction of the flow of the continental glacier that I carefully examined the northern end of the deposit to see if it was a 'tail' to a spur of the hill" (Stone, 1880, p. 301).

Our field work confirmed that both the deposits noted by Stone, and other nearby ridges, are moraines. They differ morphologically from drift tails in one or more of the following respects: (1) arcuate shape; (2) orientation transverse or oblique to ice flow; (3) presence of composite, multi-crested ridges; and (4) abrupt truncation against higher till or bedrock hills. The latter relationship is particularly noticeable on moraines 1 and 16 (Fig. 4). These high, sharp-crested moraines are very narrow in contrast to Stock Farm Mtn. and Hark Hill, and there are abrupt breaks in slope where they terminate against the sides of these hills.

The moraine segments described by Stone (1880) (14-16 and 9-11) are oriented transverse to the Androscoggin River valley. Their crests rise in elevation from as little as 720 ft (219 m) where the moraine system is breached by the river to a maximum of 880-900 ft (268-274 m) on either side of the valley. These moraines are steep-sided, generally sharp-crested, and rise abruptly 30 m or more in places. The proximal sides of moraines 9 and 16 are especially high and steep. Ridges 9 and 10 comprise at least two closely-spaced moraines with intervening kettles and short connecting ridges. Many large boulders (>1 m) are strewn along the ridges; one boulder on the proximal slope of moraine 16 is 8 m in diameter. The lithologies of these boulders are identical to rock types (discussed below) that outcrop immediately upvalley from the moraine system. Extensive search did not reveal bedrock outcrops along any of the moraine ridges shown in Figure 4.

The ridge (1) that protrudes eastward from Stock Farm Mtn. is one of the most striking segments of the Androscoggin Moraine (Fig. 5). It reaches a maximum elevation of about 1260 ft (384 m) and has a very sharp crest that stands at least 30 m above the adjacent terrain. This moraine is accessible via a logging trail that leaves an old section of U.S. Route 2, about 0.35 km west of the state line (Fig. 4). The trail follows the crest of a lower moraine (5) and then passes several other morainal deposits (2-7) before ascending the high ridge of moraine 1. The lower moraines form a cluster of nested and locally overlapping (?) ridges. Some of them are more clearly defined than they appear on the topographic map, but they are not as prominent as moraine 1.

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*Elevations are expressed in feet for ease of comparison with USGS topographic maps.*

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![Figure 5. View southwest across the Androscoggin River valley. Arrows indicate moraine ridge (no. 1 on Fig. 4) projecting eastward from Stock Farm Mtn. (middle distance) and pine-covered moraines 14-16 extending from Hark Hill (foreground).](image-url)
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Other bouldery moraines with varying degrees of topographic expression have been discovered in addition to the ones previously reported by Thompson (1983, 1986). The northernmost are the group east of Mt. Cabot (Fig. 4). The highest member of this group (moraine 19) consists of several till ridges that rise as a series of en-echelon steps to a maximum elevation of 1240 ft. (378 m), which is essentially the same elevation as the upper end of moraine 1 on Stock Farm Mtn. Moraine 21 comprises two closely spaced ridges that are not distinguished by the topographic contours.

Composition. There are few exposures of the glacial sediments in the Androscoggin Moraine system. Shallow cuts along logging trails and observations in shovel holes revealed loose, sandy, stony, light olive-gray to olive-colored diamicton. This material appears identical to ablation and resedimented facies of the typical late Wisconsinan surface till seen in crystalline-rock terrain over much of New England (Kotell and Pessl, 1985). A borrow pit exposure in moraine 18 (next to Ingalls Valley Road, Fig. 4) is badly slumped, but includes sand and gravel overlying and interbedded with bouldery till.

Five test pits were excavated by backhoe to depths of 1.8-2.4 m in order to examine the near-surface stratigraphy of the Androscoggin Moraine and collect samples for granulometric analysis and provenance studies. Four pits were located along the logging trail on the east side of Stock Farm Mtn., and one where North Road crosses the south end of moraine 15 adjacent to Hark Hill (T1-T5 in Fig. 4).

Test pit 1 showed 2.1 m of sandy, loose, bouldery diamicton. The exposed material has been thoroughly oxidized to a yellowish-brown color (10YR-5/8). It contains scattered lenses of thinly laminated, olive-gray silt and sand to 5 cm in thickness, which exhibit minor deformation. The diamicton is interpreted as till (possibly flowtill) from which most of the fines have been winnowed by meltwater.

Test pit 2 exposed 2.4 m of mostly massive, stony, sandy-silty diamicton. Pervasive oxidation is limited to the zone of modern soil development in the uppermost 1 m. Below this zone the diamicton is olive-colored (5Y-5/3). A small, deformed pod of stratified sand was seen near the bottom of the pit. The material in this exposure is chiefly till, though it may have been resedimented (as flowtill) from the adjacent ice margin.

Test pit 3 revealed a varied section in which 1.0 m of massive, pebbly to bouldery, sandy diamicton overlies 0.8 m of massive to stratified, sandy-silty diamicton interbedded with lenses of laminated silt and sand. The silt lenses are variably slumped and distorted, and contain scattered stones. The color of the nonoxidized diamicton below the modern soil zone is 5Y-5/3 (olive). We interpret the gravelly, washed upper diamicton unit to have formed as a debris flow off the ice margin in the area where moraine 2 abuts the lower end of moraine 1. The underlying stratified diamicton unit likewise was deposited in an unstable ice-marginal environment in which resedimentation occurred.

Test pit 4 exposed 1.8 m of massive, sandy-silty, olive-gray (5Y-5/2) diamicton. As in the previously described test pits, stones up to boulder size are common. The diamicton is equated with the typical surface till of the study area. It is rather compact, and water-laid sediments and collapse structures such as those seen in test pits 1-3 are lacking. The till in test pit 4 may be a different facies -- perhaps a lodgement facies -- or just a single, relatively homogeneous debris flow emplaced off the ice margin.

Test pit 5 was located in the floor of a small overgrown borrow pit in moraine 15. It showed 2.1 m of section, the upper 0.6 m of which is loose, oxidized sandy gravel. This gravel is separated by a sharp contact from the underlying nonoxidized unit, which is a well-stratified sequence consisting of three sub-units. From top to bottom, these are: (1) 10 cm of thinly laminated silt containing a few pebbles and cobbles; (2) 41 cm of moderately fissile, massive, silty-sandy, olive-gray (5Y-5/2) diamicton with scattered pebbles and cobbles; and (3) thinly laminated silt-sand with local deformation (adjacent to dropstones) and lenses of pebbly diamicton. Contacts between these sub-units are indistinct and defined mainly by the presence or absence of lamination. The gravel unit in test pit 5 is outwash, while the lower unit probably is a ponded deposit that received subaqueous debris flows (flowtills) from the adjacent glacier margin. Many large boulders occur on the moraine surface, but none were seen in the test pit.

Table 1 shows the combined percentages of silt and clay in bulk sediment samples collected from the unweathered zones in the test pits. The increase in fines from pit 1 to pit 5 is attributed to two factors. First, meltwater removed nearly all of the silt and clay from the gravelly till in moraine 1, high on the side of Stock Farm Mtn. Some of the tills in pits 2-4 retain their fines, and they locally contain interbeds of silty water-laid sediments. Second, silty beds are especially abundant in pit 5, where the morainal sediments were deposited in a ponded environment in the valley bottom.

The morphologic and stratigraphic evidence presented above demonstrate that the ridges of glacial sediment shown in Figure 4 are end moraines. The interbedded flowtills and water-laid deposits, locally contorted bedding, and angular stones scattered through the stratified units are evidence of the dynamic depositional environments and abundant meltwater that occur at the margins of temperate glaciers.

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<th>TABLE 1. WEIGHT PERCENTAGE OF SILT AND CLAY FRACTION (PERCENTAGE PASSING #200 SIEVE) IN BULK SAMPLES FROM TEST PITS 1-5</th>
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**Provenance.** Large areas of bedrock are exposed upvalley from the Androscoggin Moraine (Fig. 4). Billings and Fowler-Billings’ (1975) map of the Gorham 15-minute quadrangle shows three principal rock types in the 5-km section of the Androscoggin River valley just west of the moraine system. In decreasing order of abundance, these are: Littleton Formation (Di), consisting mostly of high-grade paragneiss, schist, and quartzite; medium-grained, gray biotite quartz diorite (qd); and Concord quartz monzonite (co), a medium-grained gray rock containing both muscovite and biotite. These rock types can be seen in road cuts along U.S. Route 2 in Shelburne. Much of what Billings and Fowler-Billings (1975) considered to be Littleton Formation in the Shelburne area was later reassigned to the Rangeley, Smalls Falls, and Madrid Formations by Hatch and Moench (1984) and Moench (1984).

Most of the boulders in the Androscoggin Moraine are the same rock types that outcrop immediately upvalley. Coarse, variably rusty, two-mica gneisses and schists of the Rangeley Formation (formerly Billings and Fowler-Billings’ Dig member of the Littleton Formation) and biotite quartz diorite are particularly common. For example, moraine 4 contains many large boulders (1–3 m), with a great concentration on the proximal side. Most of these boulders are the quartz diorite that outcrops as near as 0.25 km to the northeast. The area of outcrop of this intrusion extends across the northeast slope of Stock Farm Mtn. and is somewhat larger than shown on Billings and Fowler-Billings’ map.

Table 2 presents the results of stone counts on 100-stone samples from test pits 1, 2, 4, and 5. Clasts of the Rangeley, Smalls Falls, Madrid, and Littleton Formations were not differentiated as such, because it was not always possible to distinguish these formations in hand specimen. We have retained Billings and Fowler-Billings’ (1975) names and symbols for lithologies listed in Table 2 so that comparisons can be made with these authors’ map of the Gorham quadrangle.

Some major differences among the stone counts are apparent in Table 2. Most of the variation occurred between test pits 1 and 2 (both located in moraine 1) and pits 4 and 5, which were located nearer the center of the valley in moraines 5 and 15 (Fig. 4). It is likely that moraine 1, on Stock Farm Mtn., would have received debris from the southern margin of the Androscoggin ice tongue. This theory is supported by the relative amounts of granite pegmatite and quartz diorite in the test pits. The somewhat greater percentage of pegmatite in test pits 1 and 2 may reflect the 0.3-km wide pegmatite outcrop on the south side of the valley, 2 km west-northwest of these pits (Billings and Fowler-Billings, 1975). The abundant quartz diorite in test pits 4 and 5 probably came from the outcrop area of this distinctive rock unit northeast of Stock Farm Mtn. The sample from test pit 4, containing 59 percent quartz diorite, was taken immediately downvalley from the diorite outcrops.

Differences in percentages of other rock types among the test pits are not easily explained. The Concord quartz monzonite, for example, outcrops widely to the west and northwest of the Androscoggin Moraine, and yet is considerably more abundant in moraine 1 than in moraines 5 and 15. A noteworthy variation also exists in the combined percentages of basaltic and rhyolitic dike rocks, which total 16 percent in test pit 5, but only 0-2 percent in the other pits. The dike lithologies in test pit 5 are heterogeneous, suggesting that these rocks were not simply derived from a single nearby outcrop. Billings and Fowler-Billings (1975) compiled a map of dikes in the Gorham quadrangle. This map shows clusters of dikes on the north side of the Androscoggin River valley at Gorham and in the Berlin area. Perhaps the moraines projecting from Hark Hill (Fig. 5) are enriched in dike rocks because ice on the north side of the valley could have incorporated a greater percentage of erratics from the Gorham-Berlin dike swarms. Alternatively, the local abundances of dikes on Billings and Fowler-Billings’ map may be an artifact of the extensive outcrop in those areas (particularly at Berlin).

**Correlations.** Correlation of the Androscoggin Moraine segments on opposite sides of the valley is tenuous because of their range of elevations and the gaps caused by glacial and Holocene stream erosion. The earliest ice-margin positions are marked by moraines 8–12 and 17, which were built to maximum elevations of 850-890 ft (259-271 m) (Fig. 4). A single tongue of eastward-flowing ice extended across the Androscoggin River valley when these moraines were deposited. Moraines 6 and 7 may correlate with members of the 8-12 group, but meltwater flowing down the Connor Brook valley eroded the lower ends of these morains.
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of these ridges, destroying any former connections with the latter moraines.

The other moraines east of Stock Farm Mtn. (1-5) are younger than 6-12, and probably are contemporaneous with the moraines projecting southeast from Hark Hill (14-16). The steeply sloping crest of moraine 1 is believed to closely reflect the slope of the youngest ice margin recorded by the moraine system. This steep gradient (144 m/km), together with the apparent truncation of moraines 6-7 by 4, suggest that the ice margin thickened and readvanced slightly when the younger moraines were constructed.

The Androscoggin River valley ice lobe probably divided into three smaller sublobes as the ice receded westward past Hark Hill and Crows Nest (Fig. 4). While one sublobe deposited the moraines between Stock Farm Mtn. and Hark Hill, another extended through the gap between Hark Hill and Crows Nest, forming moraine 13 and possibly 18. The third sublobe spilled across the saddle between Crows Nest and Mt. Cabot, and deposited the 19-20-21 group. The steep profile (up to 213 m/km), bouldery crest, and upper elevation limit of moraine 19 are similar to those of moraine 1. These two moraines could have formed at the same time, when Crows Nest was beginning to emerge as a nunatak.

RELATIONSHIP OF MELTWATER CHANNELS AND DEPOSITS TO REGIONAL PATTERN OF DEGLACIATION

When glacial meltwater deposits of the northern White Mountains are considered together with those of the Mahoosuc Range-Blue Mountains in western Maine, a southeast-to-northwest pattern of deglaciation is evident. The retreat of the late Wisconsinan glacier from the study area was marked by the following events: (1) deposition of eskers and subglacial cutting of meltwater channels when glacial streams first penetrated mountain passes opening to the southeast; (2) subaerial channel cutting and minor glaciolacustrine sedimentation as the uplands were deglaciated; (3) formation of ice-contact lacustrine deposits where meltwater was ponded between the glacier margin and north- to west-facing slopes; and (4) deposition of outwash and/or glaciolacustrine sediments on valley floors when local deglaciation had reached an advanced stage. Two or more of the above events occurred in the indicated sequence in some valleys, but these events overlapped in time across the area as a whole. Gerath (1978) described similar meltwater environments in the Berlin-Gorham region.

Eskers

Englacial meltwater drainage systems developed as the last ice sheet thinned over the mountains of northern New Hampshire and western Maine. Inspection of the state surficial maps (Goldthwait et al., 1951; Thompson and Borns, 1985a) shows that several long esker systems followed hydraulic gradients leading to notches through the mountains. Examples of notches that were conduits for esker-forming streams include Crawford Notch at the south end of the Presidential Range (Goldthwait, 1940), Grafton Notch and Sawyer Notch in the northeast part of the Mahoosuc Range (Caldwell, 1975c), and other mountain passes northeast of the study area (Thompson and Borns, 1985a). Judging from the deep, narrow bedrock gorges found where some esker segments terminate at notches in western Maine, intense subglacial erosion occurred as the streams passed through these gaps. Wherever evidence of meltwater flow direction has been found in the eskers, it reveals that the streams depositing them flowed in directions ranging between south and east.

High-Elevation Meltwater Channels, Glacial Lakes, and Related Deposits

Streams issuing from the ice margin probably continued to carve channels on the floors of mountain passes as the latter sites were deglaciated. However, the relative importance of proglacial versus subglacial channel cutting is uncertain. Eskers are very small or absent at the heads of some passes such as Evans Notch (discussed below), so the meltwater channels in these places may have been cut mainly by proglacial streams. Elsewhere, hillside channels developed along the irregular margin of the thinning ice sheet. Examples were described by Gerath (1978) and Goldthwait and Mickelson (1982).

Ephemeral ice-contact lakes were ponded in topographic reentrants between the ice margin and emerging mountain slopes during deglaciation. Deposits associated with the lakes are partly deltaic, but may also include lake-bottom and glaciolfluvial sediments. Although they are mostly small and not abundant, these lake deposits indicate a generally northwestward recession of the ice margin. None of the ponded deposits known in the study area are located southeast of their corresponding spillways.

Several meltwater drainage systems are described below to illustrate the pattern of ice retreat from the uplands.

Evans Notch-Wild River Area. Shortly before deposition of the Androscoggin Moraine system, thinning and recession of glacial ice in the high mountains just south of the Androscoggin River caused meltwater to pour southward into the Saco River basin. One of the principal drainage routes was the deep bedrock gorge at Evans Notch, just east of the Maine-New Hampshire border (Fig. 6, E-1). Ponding may have occurred behind the 1400-ft (427 m) divide in this notch, but only a few small fluvial or deltaic deposits of sand and gravel were graded to the floor of the meltwater channel that crosses the divide.

 Withdrawal of the ragged ice margin from Evans Notch permitted meltwater to drain northeast along the Wild River into the Androscoggin River valley. During part of this phase, the glacier margin lay against the high ridge between the Andros-
Figure 6. Map showing meltwater drainage paths (indicated by arrows) in the uplands, including possible spillways for ice-contact glacial lakes, and outlets for glacial Lake Bethel: B - Bowman divide; C - Copperville area; E - Evans Notch-Wild River area; H - Hunters Pass; LB - glacial Lake Bethel; LC - glacial Lake Cambridge; M - Mahoosuc Range-Success Moraine area.
Deglaciation of the upper Androscoggin River valley

coggin Moraine and the Wild River. Two meltwater channels cut across this ridge just east of the state line (Fig. 6, E-2). A small ice-contact delta (?) is graded to the north end of one channel, whose elevation is about 1110 ft (338 m). The other channel is at approximately 1090 ft (332 m). The authors noted a bedrock outcrop and lag concentrate of boulders on the floor of the latter channel, which is crossed by Hastings Trail. Drainage from both outlets plunged down the steep slopes to the southeast and into the Wild River.

**Glacial Lake Cambridge.** Near the northern limit of the study area, Grafton Notch was an early outlet, at about 1550 ft (472 m), for meltwater draining south through the Mahoosuc Range (Fig. 6, LC-1). The prominent, flat-floored channel passing through the notch is located west of Route 26. A segmented esker that terminates just north of the notch was deposited when ice still filled the lowland northwest of the Mahoosucs. Some of the small, poorly exposed kames flanking the esker are high enough to be deltas graded to the Grafton Notch channel, but their identity is uncertain. Ponded meltwater draining through the notch would have constituted the earliest and highest level of glacial Lake Cambridge, which was named by Leavitt and Perkins (1935).

Recession of the ice margin caused Lake Cambridge to drain eastward through Dunn Notch (Fig. 6, LC-2), at an elevation of about 1450 ft (442 m). Deltaic topset and foreset beds were seen in a gravel pit located south of Cedar Brook on the west side of Route 26 (Fig. 1). The level of the topset/foreset contact in this delta corresponds to the Dunn Notch spillway.

Further deglaciation may have resulted in eastward drainage of Lake Cambridge across the C Pond Moraine (Fig. 1, CPM) and down through Sawyer Notch. The elevation of the lowest possible spillway across the moraine is likewise approximately 1450 ft (442 m). This spillway could not have drained the lake for very long; slight additional recession of the ice from the Upton area opened lower outlets at 1270-1330 ft (387-405 m) into the Androscoggin River valley northwest of Hampshire Hills (Fig. 1). The demise of Lake Cambridge presumably occurred when it emptied directly into the head of the Androscoggin River near Errol. Additional field work is needed to verify the lake spillways and to determine the regional significance of the C Pond Moraine (originally reported by Caldwell, 1974).

**Success and Copperville Moraines.** Progressive northwestern recession of the ice margin is also recorded by meltwater channels and deposits in the part of the Mahoosuc Range near Berlin. Initial drainage occurred at high levels soon after the peaks emerged from the ice. Leavitt and Perkins (1935) and Gerath (1978) described Mahoosuc Notch (Fig. 6, M-1) as one of the earliest drainage channels. Its elevation is about 2470 ft (753 m).

Gerath (1978) reconstructed the deglaciation of the New Hampshire portion of the Mahoosuc Range. He found that retreat of the ice margin down the proximal side of the range allowed meltwater to drain southwestward into the Androscoggin River valley, carving a series of channels that are lower to the northwest (Fig. 6, M-2,3,4). Gerath discovered thick ice-contact deposits in the township of Success and called them the "Success Moraine" (Fig. 1, SM). He described this moraine as "a complex mass of outwash and glaciolacustrine sediment which formed in a depositional environment confined between a broad, retreating Laurentide ice front and the north slopes of the Mahoosuc Range" (Gerath et al., 1985, p. 23).

The Success Moraine reaches elevations of slightly over 1600 ft (488 m). On topographic maps it does not show obvious ice-marginal ridges, and sedimentation in ponded water played a major role in its formation. The moraine is a lacustrine ice-contact morphosequence in the terminology of Koteff and Pessl (1981). Gerath (1978) described an exposure that indicates that the ice margin advanced over the proximal side of the Success Moraine. It shows 6 m of till overlying 12 m of deformed glaciolacustrine sediments. Gerath concluded that the moraine was deposited at the terminus of an active ice tongue that persisted in the Androscoggin River valley, but that the lower meltwater channels a short distance to the northwest formed as the ice tongue stagnated. Gerath noted similarities between these channels and modern analogs adjacent to stagnating ice in Alaska.

The Copperville Moraine (Fig. 1, CM), also named and first described by Gerath (1978), is located northwest of Berlin in what he called the "Copperville through-valley." This valley crosses the divide at about 1070 ft (326 m) between the Connecticut and Androscoggin River basins. The Copperville Moraine is a small transverse ridge of stratified sand and gravel, situated approximately 3.5 km northwest of the divide. Gerath (1978) interpreted it as being the collapsed head of a fluvial ice-contact morphosequence that was built across the divide and into the Dead River valley to the southeast. The moraine was thought to have been deposited adjacent to active ice flowing from the Connecticut River basin and is somewhat lower and younger than the Success Moraine. Although it was identified as a fluvial deposit, some meltwater ponding probably occurred on the proximal side of the Connecticut-Androscoggin divide (Fig. 6, C).

**Glacial Lakes South and West of Gorham.** Field evidence suggests that meltwater was ponded in the lower part of the Peabody River valley (Fig. 1). About 2.5 km south of Gorham village, next to where Route 16 crosses the river, a cut bank shows a diamicton unit (probably till) overlying rhythmically bedded lacustrine sediments. This exposure was briefly mentioned by Crosby (1934) and subsequently described by Gosselin (1971), Goldthwait (1972), and Davis et al. (1988). It is believed to represent overriding of glacial-lake beds by an ice tongue advancing southward from the Androscoggin River valley. Until more work is done on this section, several interpretations must be considered. The deposits seen here may have formed during either the advance or recession of the late Wisconsinan ice sheet; and it is also possible that the lacustrine sediments predate the last glaciation.

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Other features besides the Copperville Moraine demonstrate that meltwater from ice in the Connecticut River basin spilled eastward into the Androscoggin River valley. One of the higher meltwater channels is Ice Gulch (also known as Icy Gorge), a 40-m deep gorge illustrated by Goldthwait and Mickelson (1982). This channel heads at the 2490-ft (759 m) gap called Hunters Pass (Fig. 6, H). It drained meltwater from ice northwest of the Crescent Range.

West of Gorham village, there is a deep east-west glacial trough between the Presidential Range to the south and the Crescent Range to the north. Known as "Randolph Valley" (Goldthwait, 1940), it straddles the drainage divide at Bowman, whose elevation is about 1500 ft (457 m) (Fig. 6, B). This divide separates the Moose River, which joins the Androscoggin at Gorham, from Israel River, which flows westward to the Connecticut River. Randolph Valley channeled meltwater into the Androscoggin River valley from a broad ice lobe in the upper Connecticut River basin.

Gerath (1978) found that meltwater draining through Randolph Valley deposited the Moose Brook delta in a small glacial lake at Gorham. Gerath also mapped a younger outwash fan from the same source in Moose Valley. These deposits indicate that glacial streams continued to transport sediment eastward through Randolph Valley until the Gorham area was essentially ice-free. Gerath (1978) concluded that the upper Connecticut River basin probably held the latest active ice in the region.

Goldthwait et al. (1951) and Goldthwait and Mickelson (1982) mapped a series of "ice-contact deposits" west of the Bowman divide. Goldthwait (1940) described these deposits as gravel hummocks that are lower in elevation to the east. Descending sets of meltwater channels occur on both sides of Randolph Valley at the proximal end of this series of kames (Goldthwait and Mickelson, 1982). Lougee (1939) published a photograph of a pothole in the bedrock floor of one channel. Perhaps the kames are chiefly fluvial deposits, in which case they may be contemporaneous with the outwash carried down the Moose River valley to Gorham. However, the present authors saw delta foreset beds in some of the deposits west of Bowman, and Lougee (1939) considered this to be an area of deltaic sedimentation. Both fluvial and lacustrine environments may have existed in turn as the Israel River valley was deglaciated.

Other Ice-Marginal Lakes. In addition to the examples discussed above, there probably are other places where meltwater was ponded between mountain slopes and the ice margin. Lougee (1939) compiled a map showing an extensive network of lakes that he thought existed along the glacier margin as it receded northwestward from the upper Connecticut River basin in New Hampshire. Evidence for some of these water bodies, such as glacial Lake Ammonoosuc, has been discussed in the literature (Goldthwait, 1916; Lougee, 1940). Certain other lakes proposed by Lougee (1939) are not supported by field evidence (R. P. Goldthwait, pers. comm., 1987). Nevertheless, the distribution of ice-contact deposits and associated meltwater channels in the areas described above indicate a consistent pattern of northward retreat of an irregular ice margin from the mountains of northern New Hampshire and adjacent Maine.

Meltwater Deposits along the Androscoggin River

The upper Androscoggin River valley contains both glaciofluvial and glaciolacustrine deposits. The fluvial deposits include scattered esker segments, but mostly comprise dissected outwash which is locally kettled and collapsed from deposition over dead ice. Gerath (1978) has described the meltwater deposits from the state line upstream to Milan, New Hampshire, so this area will not be discussed further.

Geologic mapping in the Androscoggin River valley of western Maine has not progressed to the point where we can reconstruct the sequence of meltwater deposits in detail. However, gravel-aquifer studies by Williams et al. (1987) suggest the presence of a single long outwash sequence extending from Gorham east to Bethel. The elevations of the dissected outwash remnants become gradually lower downvalley, from about 840 ft (256 m) at Gorham to 700 ft (213 m) at Bethel. There is an overall decrease in particle size and degree of ice-contact collapse along this section of the valley. Bouldery, poorly-sorted gravel in pits at Gorham contrasts with the well-stratified sand and pebble gravel of the Bethel area. In places, however, a greater abundance of kettles, together with kames or ice-channel fillings at higher elevations, may indicate the presence of older sequence heads that were modified by the later drainage from New Hampshire.

The meltwater from the upper Androscoggin River valley emptied into a glacial lake at Bethel (Fig. 1). Evidence for this lake includes a delta at West Bethel. A topset/foreset contact in the delta was prominently exposed along the south bank of the Androscoggin River following the spring flood of 1987. The elevation of the contact (which indicates the former lake level) is approximately 690 ft (210 m). Exposures of delta topset and foreset beds were also seen in borrow-pit exposures, and lake-bottom sediments were penetrated by an observation well (well LB-2) is a channel at about 690 ft (210 m), which carried water into the northeast corner of Songo Pond.

Deglaciation of the Bethel lowland eventually allowed Lake Bethel to drain northeast along the Androscoggin River valley. The last impoundment of the lake may have occurred behind a till barrier that probably blocked a narrow segment of the valley in Newry, and which has since been incised by the modern river (Fig. 6, LB-4).
DISCUSSION

There has been much controversy as to whether the White Mountains were deglaciated mainly by regional stagnation and downwastage of the ice, or by progressive retreat of an active ice margin. Johnson (1941) attempted to reduce the confusion by pointing out that "normal ice retreat" and "down-wasting" are not mutually exclusive processes. Thinning by ablation of the terminal region of a terrestrial ice sheet can result in recession of the glacier margin regardless of whether the ice is stagnant or actively flowing. Moreover, the configuration of the margin would be irregular in a high-relief terrain such as the White Mountains, even if deglaciation occurred by recession of active ice.

Goldthwait and Mickelson (1982) advocated the concept of early and pervasive glacial stagnation as the last ice sheet thinned over the White Mountains. They also noted that there was a northwestward progression in the uncovering of valleys. Gerath et al. (1985) discussed the importance of topography in controlling the dynamics of the thinning ice sheet and concluded that the Copperville and Success Moraines formed at the edge of ice streams that persisted in valleys open to the northwest. Meltwater channels and ice-contact deposits like those described in the previous section of this report commonly form in association with stagnant ice (Goldthwait and Mickelson, 1982), and stagnation presumably did occur in valleys that were cut off from ice flow. However, our work on the Androscoggin Moraine supports the conclusion that active ice was present in the northeastern White Mountains until an advanced stage of local deglaciation, when the Presidential and Mahoosuc Ranges had largely emerged from the late Wisconsinan ice sheet. The position of the Androscoggin Moraine suggests that it is slightly older than the Success Moraine, but certainly younger than the meltwater drainage through higher channels to the south. Thus we infer that an irregular but still-active ice margin retreated across the Maine mountains south of the Androscoggin River just prior to deposition of the moraine. Local stagnation eventually occurred as ice masses separated over the mountains, especially in the lee of the Mahoosuc Range, but ice flow continued in the upper Androscoggin River basin.

The limited striation evidence indicates that the ice which deposited the Androscoggin Moraine and stratified moraines of the Berlin-Gorham area did not originate in the lowland at the head of the Androscoggin River. Instead, it is believed to have flowed from the west and northwest as part of an ice lobe in the upper Connecticut River basin. This lobe extended through the Copperville and Randolph Valleys and deposited the moraines described above. Gerath (1978) found evidence that ice persisted in the Connecticut River basin, shedding debris and meltwater into the Androscoggin River valley until the latter area was almost totally ice-free.

No field evidence has been discovered from which to reconstruct the ice-surface profile across the mountains of northern New Hampshire when the Androscoggin Moraine was deposited. However, J. L. Fastook (Quaternary Studies Institute, Univ. Maine) developed a preliminary computer model of the ice profile extending from the moraine north-northwestward across the mountains for a distance of 85 km. Several combinations of ice-floor elevations and glaciological conditions were tested using Fastook's model, and each of the reconstructed ice profiles was high enough to cover all (or nearly all) of the mountainous terrain upglacier from the Androscoggin Moraine. Therefore, it seems likely that the moraine was deposited by ice flowing out of the upper Connecticut Valley region, though how far to the north this flow originated is uncertain.

The rarity of moraines in the White Mountains and western Maine leads us to question why the Androscoggin Moraine system formed in the first place. It could be the result of a local glacial readvance, but there is no independent evidence to support this theory. Nor is there any reason why a readvance capable of depositing such a large moraine cluster would not be recorded elsewhere in the study area.

Examination of the regional topography shown on the Lewiston 1:250,000 quadrangle reveals another possible explanation for the Androscoggin Moraine. The highest part of the White Mountains extends from the Franconia Range northeast to the Presidents and hence to the Mahoosuc Range and Blue Mountains in Maine. Together, these ranges form a continuous northeast-trending mountain chain that obstructed the southeast flow of the late Wisconsinan ice sheet. The ice was funneled into narrow valleys that penetrate the mountains and eroded deep troughs such as Franconia Notch and Crawford Notch. Along a broad section of the mountain front including the Presidents and Mahoosucs, ice flow was directed into the constricted segment of the Androscoggin River valley east of Gorham. The conditions were similar to drainage of modern ice caps by outlet glaciers (Embleton and King, 1975). Accelerated flow occurred in the narrow ice stream extending down the Androscoggin River valley. A balanced condition was attained for an uncertain length of time, during which the ice margin stood at the Maine-New Hampshire border and deposited the Androscoggin Moraine. The large boulders and considerable volume of sediment in the moraine ridges attest to the erosion caused by this ice stream.

It remains to determine the age of the ice-marginal deposits discussed here and the relationship of ice retreat in the study area to the deglaciation history of northern New England and adjacent Quebec. Davis and Jacobson's (1985) synthesis of radiocarbon dates shows that the receding late Wisconsinan ice margin reached the inland marine limit in southwestern Maine by about 13,000 yr B.P. A reconnaissance of ice-contact deposits and meltwater channels indicates that the glacier margin then receded generally northwest from the marine limit to the White Mountains (W. B. Thompson, unpublished data). On the basis of these inferences, the upper Androscoggin Valley region may have been deglaciated sometime after 13,000 yr B.P. This chronology is compatible with Gerath's (1978) estimate that the Gorham area became ice-free between 12,600 and 12,100 yr B.P., but probably is too young for reasons discussed below.
If the study area was deglaciated after 13,000 yr B.P., there is a question as to whether the recessional deposits were formed by the main Laurentide ice sheet or a residual Appalachian ice mass that existed along the Maine-Quebec border in late-glacial time (Lowell and Kite, 1988). This ice mass began to detach from the Laurentide ice sheet prior to 13,000 yr B.P., as a marine calving bay encroached upon the ice stream flowing down the lower St. Lawrence Valley (Dyke and Prest, 1987). The Champlain Sea expanded into the St. Lawrence Lowland southwest of Quebec City no later than about 12,000 yr B.P., by which time the separation of the Appalachian ice mass was complete (Dyke and Prest, 1987; Parent and Occhietti, 1988). However, it is doubtful that this remnant ice extended far enough to the southwest to affect northern New Hampshire or adjacent western Maine. Numerous end moraines and other ice-contact deposits in the Sherbrooke-Asbestos area of Quebec indicate a progressive northwestward recession of the Laurentide ice margin from the northern tip of New Hampshire (Parent and Occhietti, 1988).

A difficulty with proposing ice retreat from the study area after 13,000 yr B.P. would be the lack of agreement with the deglaciation model of Davis and Jacobson (1985). These authors depicted the White Mountains and Mahoosuc Range as protruding from the thinning ice sheet by 14,000 yr B.P., and deglaciation of northernmost New Hampshire and the Boundary Mountains (on the Maine-Quebec border) by 13,000 yr B.P. Dyke and Prest (1987) likewise inferred that the ice margin lay a short distance north of the Quebec-New Hampshire border at 13,000 yr B.P. This sequence favors construction of the ice-marginal deposits in the northern White Mountains by the Laurentide ice sheet at about 14,000 yr B.P. Davis and Jacobson’s model also shows a broad area of remnant ice over southwestern Maine that was cut off from the ice sheet northwest of the emerging mountains. No unequivocal field evidence has been found to support the existence of such a large remnant ice mass in this part of Maine, although striations at several localities south of the Androscoggin River indicate a late shift in glacial flow to the south and even to the south-southwest (W. B. Thompson, unpublished data).

We conclude that the upper Androscoggin River valley and adjacent mountains were deglaciated by the Laurentide ice sheet before 13,000 yr B.P., and perhaps as early as 14,000 yr B.P. This is based on Davis and Jacobson’s (1985) analysis of New England radiocarbon dates, the ice-margin chronology proposed by Dyke and Prest (1987), and the synthesis by Parent and Occhietti (1988) of studies in southeastern Quebec. A possible discrepancy remains between deglaciation of the study area by 14,000 yr B.P. and a radiocarbon date indicating the presence of an active ice margin at Kennebunk in southwestern coastal Maine as recently as 13,200 yr B.P. (Smith, 1985). There are very few dates on actual ice-margin positions in northern New England. A more consistent pattern of ice retreat should become apparent as additional dates are obtained and combined with the results of intensive field studies.

SUMMARY

Early investigators of the deglacial history of the White Mountains often described small parts of this region, such as the Bethlehem Moraine or the deposits of glacial Lake Ammonoosuc. Some of the ensuing controversies over the style of deglaciation narrowly focused on these deposits, with the result that the overall history of retreat of the late Wisconsinan ice sheet from the mountains remained poorly documented. Arguments about the presence or absence of local glaciers were followed by equally intense debate concerning the mode of ice recession.

Recent workers such as Goldthwait and Mickelson (1982) and Gerath et al. (1985) have taken a broad perspective on the deglacial history of the White Mountains and have generally favored regional stagnation and downwasting as the principal means by which the ice disappeared. However, a study of the Berlin-Gorham region by Gerath (1978) and the present authors’ work on the Androscoggin Moraine suggest that active ice flow persisted in the upper Androscoggin and Connecticut River basins until the Presidential, Carter, and Mahoosuc Ranges were largely deglaciated. The Androscoggin Moraine resulted from topographic funneling of ice flow and development of a rapidly flowing ice stream in the constricted east-west segment of the Androscoggin River valley. Gerath (1978) and Gerath et al. (1985) concluded that moraines in the Berlin-Gorham area likewise were deposited by ice streams. However, the latter moraines are morphosequences of stratified drift, in contrast to the sharp-crested ridges of boudery till with interbedded meltwater deposits that comprise the Androscoggin Moraine.

Meltwater channels and deposits of ice-contact stratified drift reveal that systematic northwestward recession of the continental ice margin occurred throughout the northern White Mountains and adjacent western Maine. Retreat of an active ice margin was the likely mode of deglaciation in this region, especially on the northwest side of the White Mountains and Mahoosuc Range, though there was widespread stagnation of detached ice masses in the lee of the mountains.

Deglaciation of the study area probably occurred by retreat of the Laurentide ice sheet as early as 14,000 yr B.P. Borns (1985) similarly concluded that the nearby Frontier Moraine on the Maine-Quebec border was deposited along the Laurentide ice margin prior to 13,000 yr B.P. A contrasting model would deglaciate the upper Androscoggin River valley by progressive northwestward retreat of an ice margin that is thought to have lain near the Maine coast as recently as 13,200 yr B.P. This interpretation suggests that the Androscoggin Moraine and other ice-marginal deposits in the study area were deposited after 13,000 yr B.P. and may have been formed by a residual Appalachian ice mass. The latter model is problematic because field studies in southeastern Quebec have shown that remnant Appalachian ice on the Maine-Quebec border did not extend far enough to the southwest to affect the White Mountains, and that the Laurentide ice margin retreated progressively northwestward.
from the northern tip of New Hampshire (Parent and Occhietti, 1988). Moreover, Davis and Jacobson's (1985) interpretation of New England radiocarbon dates and the paleogeographic reconstructions by Dyke and Prest (1987) indicate separation of the late Wisconsinan ice sheet over the White Mountains and Mahoosuc Range by 14,000 yr B.P., and total deglaciation of our study area by 13,000 yr B.P.

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