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NO EVIDENCE FOR POST-ICESHEET CIRQUE GLACIATION IN NEW ENGLAND

RICHARD B. WAITT* AND P. THOMPSON DAVIS**

ABSTRACT. New field data and scrutiny of past reports reveal no firm evidence that local cirque glaciers were rejuvenated after late Wisconsin continental ice disappeared from the northern Green Mountains of Vermont, Mount Katahdin in Maine, and the northern Presidential Range of New Hampshire. The sharpness of some cirques and arêtes in New England is not reliable evidence that the cirques were last occupied by local glaciers, for even sharper cirques and arêtes have survived overriding by icesheets in other regions. Alleged low-altitude cirques in the Green Mountains seem actually to be stream-eroded valleys modified by the overriding icesheet. Abundant erratics within cirque-floor drift in all three areas suggest that the last identifiable glacier to occupy the highland cirques and adjacent valleys was the melting icesheet. The general absence of arcuate end moraines in and downslope from cirques in all three areas suggests that local cirque glaciers did not rejuvenate after icesheet deglaciation. Bouldery ridges near some cirques seem to be deposits of wasting icesheet tongues and of later rapid slope processes. Much diamict that may be mistaken for alpine till in the Green Mountains and Presidential Range consists of debris-flow and alluvial deposits derived from unvegetated icesheet drift just after deglaciation.

INTRODUCTION

Whether local glaciers were reborn after the most recent (late Wisconsin) icesheet disappeared from high mountains in New England (fig. 1) has been an enduring argument. At the Presidential Range in New Hampshire, Thompson (1960, 1961) and Bradley (1981) argue for post-icesheet local glaciers, but Goldthwait (1970a) and Fowler (1984) against them. At Mount Katahdin in Maine, Caldwell (1972, 1980) argues for a rebirth of local ice, but Davis (1976, 1978, 1983) demurs. Over the years strong arguments for post-icesheet local glaciers at various high-mountain areas in New England and adjacent areas (Tarr, 1900; Rich, 1906, 1935; Johnson, 1917, 1933; Antevs, 1932; Flint, Desmorest, and Washburn, 1942; Flint, 1951; Thompson, 1960, 1961; Caldwell, 1966, 1972, 1980; Wagner, 1970; Craft, 1979; Bradley, 1981) are balanced by various arguments against post-icesheet local glaciers (Fairchild, 1913; J.W. Goldthwait, 1913, 1916; R.P. Goldthwait, 1939, 1940, 1970a; Borns and Calkin, 1977). Table 1 frames the nature of arguments on both sides. Northeast of New England, highlands in Newfoundland, Nova Scotia, New Brunswick, and Gaspé give clear and uncontested evidence of late-glacial radial outflow (Grant, 1969, 1977; Brookes, 1970; Waitt, 1981; David and Lebuis, 1985). But those areas

* U.S. Geological Survey, 5400 MacArthur Boulevard, Vancouver, Washington 98661

** Department of Natural Sciences, Bentley College, Waltham, Massachusetts 02254

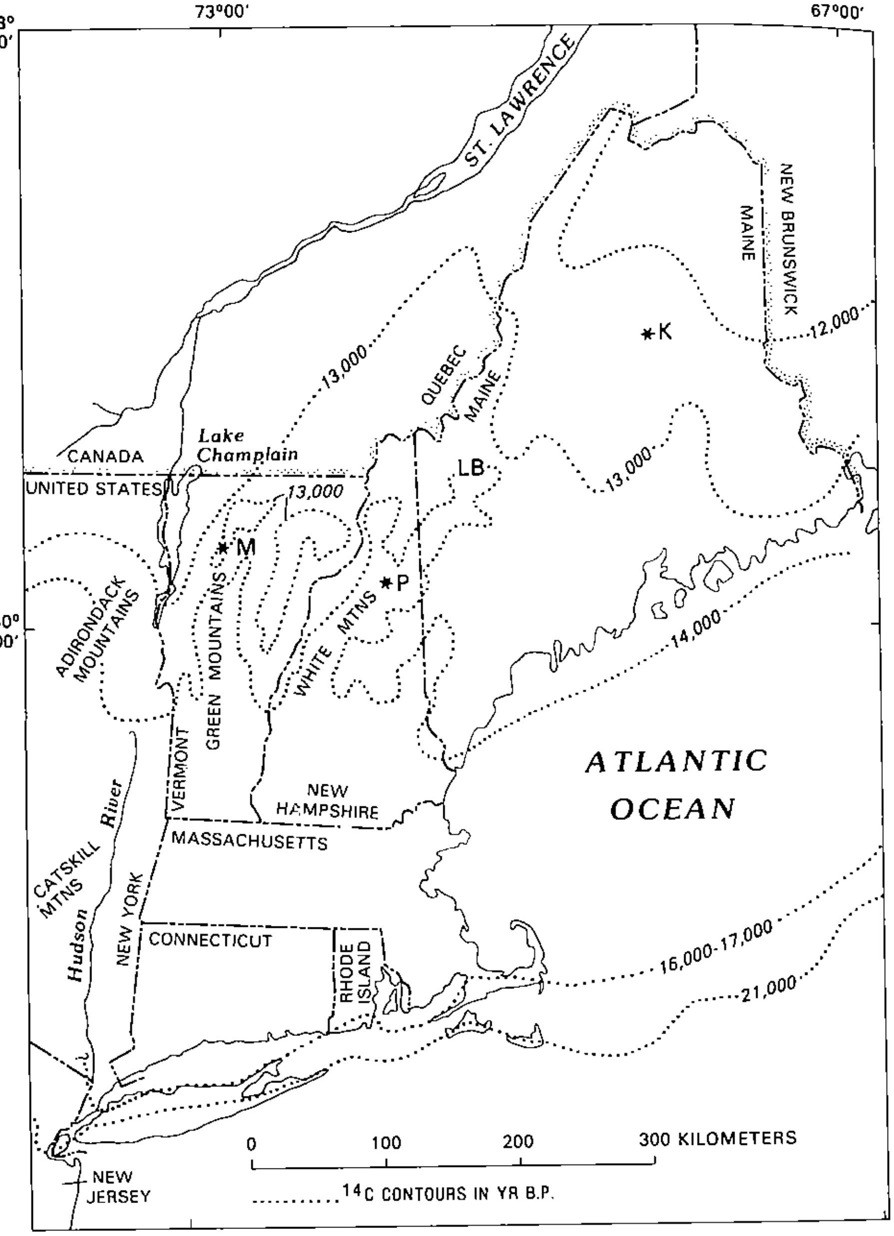
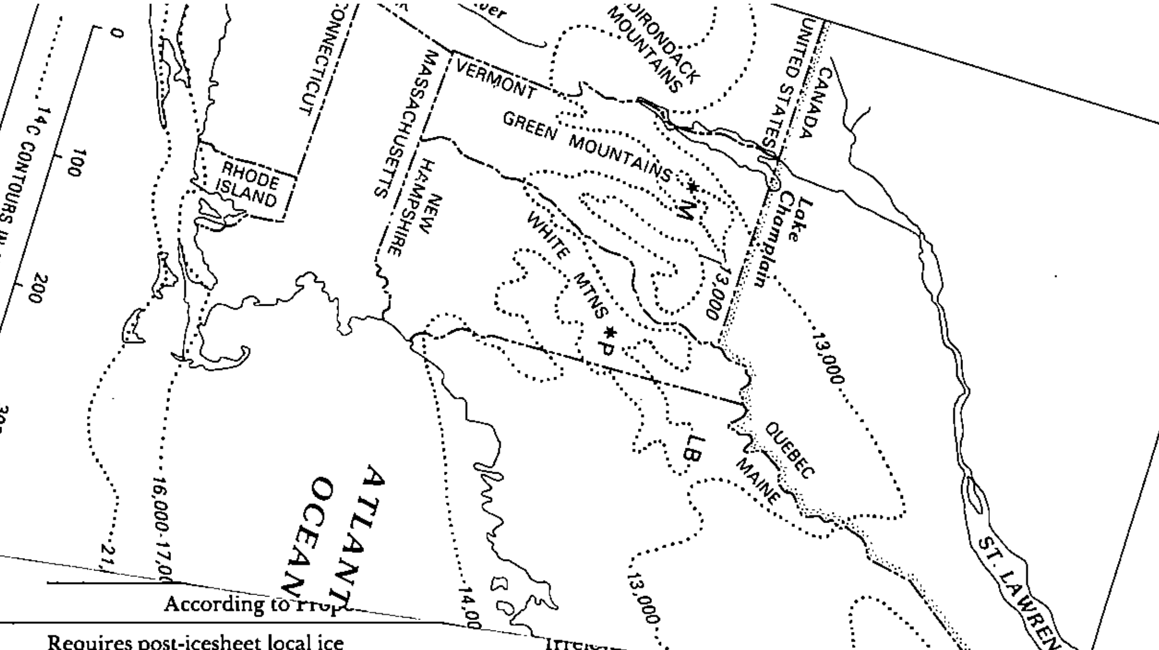


Fig. 1. Map showing recessional icesheet margins (dashed lines) and high-mountain areas in New England. Areas discussed in this report are: M, Mount Mansfield areas; P, northern Presidential Range; K, Mount Katahdin. Longfellow and Boundary Mountains (LB) in west-central Maine discussed by Borns and Calkin (1977). Dashed contours show recessional icesheet margins from Denton and Hughes (1981, fig. 2-2) as modified by Davis and Jacobson (1985, fig. 9). Other recent reconstructions of ice-margin recession show different shapes and values of these isochrons (Bonnichsen, Jacobson, and Bourns, 1985, fig. 1; Hughes and others, 1985, fig. 4; Smith, 1985, fig. 10), but these differences are not important to this report.



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Evidence	According to Prop...	
Steep cirques and sharp arêtes	Requires post-icesheet local ice (K 1,7,15,17,28,35,45; P 15; G 24; N 2,9)	Irre... basins, or icesheet... forms was minor (G 18; P 21; K ...)
Upcirque-pointing roches moutonnées on cirque floors		Indicates icesheet was last (P 21; K 29)
Icesheet erratics in cirque-floor drift	Permits late cirque glaciers that did not completely remove icesheet drift (K 15,28)	Indicates icesheet was last (P 3,5,11,12,21,22,30,39; M 31)
Moraines in or downvalley of cirques	Are analogous to pure alpine features and thus demand post-icesheet local glaciers (K 1,7,15,17,28,35,45; G 24,25,27; P 7,15,38; N 2,6,8,9,33)	Irrelevant: moraines are of icesheet or postglacial origin (P 10; K 29,32,34,41; G 16,18,19,26,27; M 31)
Apparent absence of moraines in and downvalley of cirques	Moraines are obscure because they are small, subdued by postglacial processes, or thickly forested (P 6,15,37,38; K 15; N 4)	Negative evidence for lack of post-icesheet local ice (P 5,11,21)
Till-like drift downvalley from cirques rich in lithologies derived from upslope	Indicates local glaciers (C 13; N 33; P 37,38)	Irrelevant: deposited by postglacial processes (P 22,40,42,43)
Waterlaid drift downstream from cirques	Too voluminous not to imply source from alpine glaciers (G 24)	Irrelevant: deposited by receding icesheet (G 16,18,19,26)
General radial patterns of striae, drift, moraines	Various highlands and coastal lowlands experienced late-glacial radial flow of local ice (C 13; R 14; K 46)	Such positive evidence exists in maritime areas influenced by marine calving of orphaned icesheet scraps (C 20,23,36,44)

*P, Presidential Range; K, Mount Katahdin; G, Green Mountains; M, west-central Maine; N, New York (Adirondacks, Catskills); C, Canada (Gaspe, Nova Scotia, Newfoundland); R, regional

†1, Tarr (1900); 2, Rich (1906); 3, J.W. Goldthwait (1913); 4, Fairchild (1913); 5, J.W. Goldthwait (1916); 6, Johnson (1917); 7, Antevs (1932); 8, Johnson (1933); 9, Rich (1935); 10, J.W. Goldthwait (1938); 11, R.P. Goldthwait (1939); 12, R.P. Goldthwait (1940); 13, Flint, Desmorest, and Washburn (1942); 14, Flint (1951); 15, Thompson (1960–61); 16, Stewart (1961); 17, Caldwell (1966); 18, Connally (1968); 19, Stewart and MacClintock (1969); 20, Grant (1969); 21, Goldthwait (1970a); 22, Goldthwait (1970b); 23, Brookes (1970); 24, Wagner (1970); 25, Wagner (1971); 26, Stewart (1971); 27, Connally (1971); 28, Caldwell (1972); 29, Davis (1976); 30, Goldthwait (1976); 31, Borns and Calkin (1977); 32, Davis (1978); 33, Craft (1979); 34, Davis and Davis (1980); 35, Caldwell (1980); 36, Waitt (1981); 37, Bradley (1981); 38, Bradley (1982); 39, Goldthwait and Mickelson (1982); 40, Gerath and Fowler (1982); 41, Davis (1983); 42, Fowler (1984); 43, Gerath, Fowler, and Hazelton (1985); 44, David and LeBuis (1985); 45, Caldwell, Hanson, and Thompson (1985); 46, Caldwell and Hanson (1986).

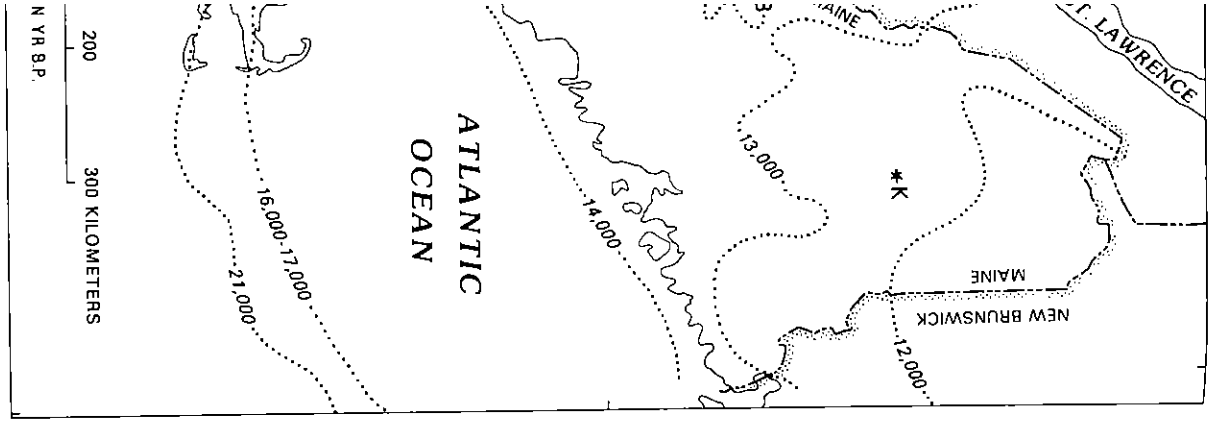

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TABLE 1

Summary of previous views on post-icesheet mountain glaciation in uplands in New England and adjacent areas

Evidence	Implications for Post-icesheet Mountain Glaciation*† According to Proponents	According to Opponents
Steep cirques and sharp arêtes	Requires post-icesheet local ice (K 1,7,15,17,28,35,45; P 15; G 24; N 2,9)	Irrelevant because the icesheet eroded out the basins, or icesheet modification of older alpine forms was minor (G 18; P 21; K 29)
Upcirque-pointing roches moutonnées on cirque floors		Indicates icesheet was last (P 21; K 29)
Icesheet erratics in cirque-floor drift	Permits late cirque glaciers that did not completely remove icesheet drift (K 15,28)	Indicates icesheet was last (P 3,5,11,12,21,22,30,39; M 31)
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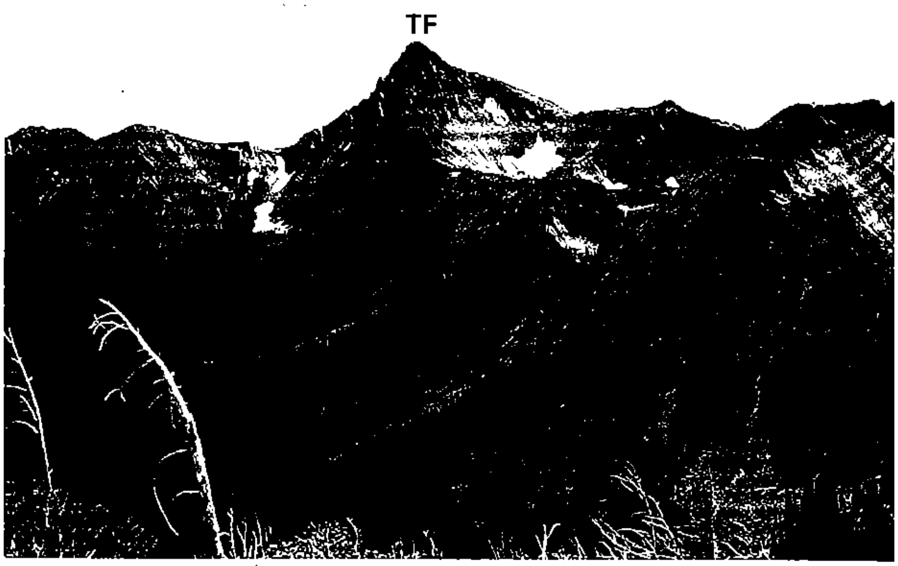

Fig. 2. Cirques overridden by Cordilleran icesheet along northern crest of Cascade Range, Wash. (A) View south from Hopkins Pass. Cirque floors as well as summit ridges in area are littered with granitic erratics denoting last occupancy by Cordilleran ice (Waitt, 1977, fig. 5); TF, Three Fools Peak.

are surrounded by ocean; late-glacial calving in deepening seawater in the Laurentian channel and St. Lawrence estuary segregated icesheet remnants there from the main Laurentide ice mass to the north (Hughes and others, 1985).

We present a synthesis of published accounts and new field evidence and conclude that post-icesheet mountain glaciation either did not occur in New England at all or was so short lived and areally restricted that it left few or no diagnostic features. Responsibility for fieldwork is: Cascade Range (Washington) and Green Mountains (Vermont) by Waitt, Mount Katahdin (Maine) by Davis, Presidential Range (New Hampshire) by both of us.

TEMPERATE ICESHEET IN HIGH-RELIEF TERRAIN

Temperate icesheets produce features in high-relief terrain that may resemble mountain-glacial effects. Mountain-glacial cirques overridden by the thick Cordilleran icesheet in the North Cascade Range, Wash. (fig. 2) (Waitt, ms and 1975, 1977) are at least as sharp as any that have figured in the New England debate. Cordilleran ice widened some low-altitude Cascade valleys into straight, U-shaped "glacial" troughs with truncated spurs. During deglaciation, tongues of Cordilleran ice

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(B) View southeast of ragged headwall and arête of cirque at Three Fools Peak. These overridden cirques and arêtes are at least as sharp and developed as those in New England on the north and east sides of the Presidential Range and Mount Katahdin.

receding up valleys built arcuate moraines like those typically formed by mountain glaciers. Thus in a high-relief terrain, invasionary icesheets have left erosional topography carved by earlier mountain glaciers more or less intact and have produced erosional and depositional features like those of mountain glaciers.

The presence of far-traveled erratics in broad-spread drift is definitive evidence of icesheet glaciation. But a succeeding alpine glacier may exist for too short a time to remove all older material from a cirque, or icesheet erratics initially deposited on a valley side may move by postglacial mass wasting to a cirque floor last occupied by local ice. Both possibilities must be discounted before an erratic-containing cirque can be proven to have been last occupied by an invasionary icesheet. But conversely, one must have definitive proof in *uniquely* alpine deposits before post-icesheet mountain glaciation is more than an *ad-hoc* idea.

MASS-WASTING AND ALLUVIATION IN HIGH-RELIEF TERRAIN

Postglacial debris flows and floods may deposit coarse diamicts on valley sides and floors that resemble till, especially if derived partly from till upslope. These mass-wastage processes probably were orders of magnitude more rapid just after deglaciation and before forestation than they have been in historic time (Church and Ryder, 1972). Several areas in the Cascade Range, Wash. give evidence that unvegetated slopes mantled with loose glacial deposits rapidly shed large volumes of this debris to the lower valley sides and floors just after deglaciation. On the floors of valleys from which late-Wisconsin alpine glaciers receded east of Glacier Peak volcano, some alluvial and debris-flow fans reached nearly their present volumes before the 11,250-yr-BP Glacier Peak tephra fell. Much or most of the existing alluvium and mass-wastage debris therefore accumulated just after deglaciation, probably before forestation. Near Mount St. Helens the landscape devegetated by the cataclysmic eruption of May 1980 has changed at extremely high rates since 1980. Not only have valley-floor channels been cut tens of meters wide and deep and kilometers long, but scores of debris flows have descended slopes and carried to valley floors enormous logs, boulders, and voluminous finer debris.

A philosophy of long-term gradualism that influenced earlier interpretations of alluvial deposits (Wolman and Miller, 1960) is being challenged by recognition of the dominating effects of brief, high-energy events (Beaty, 1974; Kochel and Baker, 1982; Wells and Harvey, 1987). During exceptional storms even currently forested slopes may deliver coarse material, including large boulders, to mountain-valley floors. Unusual heavy rains have recently triggered hundreds of damaging bouldery alluvial and debris flows to valley floors in the Basin and Range, Washington Cascade Range, northwest England, and many other areas (Beaty, 1963; Gallino and Pierson, 1985; Wells and Harvey, 1987). Catastrophic streamflows and debris flows move boulders over gentle slopes at the bases of many mountains, characteristically leaving downslope-trending levees (Blackwelder, 1928; Krumbein, 1942; Sharp, 1942; Sharp and Nobles, 1953; Beaty, 1963, 1970; Johnson, 1970, chaps. 12 and 13; Rodine and Johnson, 1976; Bradley and Mears, 1980; Costa and Jarrett, 1981). In the White Mountains of New Hampshire, heavy rain caused bouldery debris flows to reach the valley floors in 1826, 1927, and 1936 (Morse, 1966, chap. 8; Goldthwait, 1970b). Such deposits, which may include reworked glacial drift or contain locally derived boulders as large as 5 m, generally have not been distinguished from primary late Wisconsin icesheet deposits in northern New England. Deposits of debris avalanche, debris flow, and stream may be

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not only texturally similar to each other but also easily mistaken for glacial drift (Crandell and Waldron, 1956; Beaty, 1961; Crandell, 1971; Porter and Orombelli, 1980a and b, 1982).

Deposits we consider to be till in the Cascade Range and the New England uplands vary with bedrock lithology but generally are nearly massive stony diamicts with a granular muddy-sand matrix. Clasts range from angular to rounded; some are faceted with flat to slightly curved faces, a few with diagnostic snub-nosed pentagonal shape (Flint, 1971, fig. 7–13). Some stones are striated parallel to their longest dimension. These and other characteristics of texture and clast shape of tills have been amply described (Flint, 1971, p. 152-169; Goldthwait, 1971; Boulton, 1971; Drake, 1971).

While mapping formerly glaciated valleys in the Cascade Range, Waitt (1982, 1987) found many valleys to be discontinuously blanketed by diamicts that contain glacially faceted or striated stones. But where visible in section, many of these diamicts are noncompact, contain abundant angular local-rock fragments, include prominent water-winnowed layers, or show geomorphic evidence such as downslope-trending levees that they are catastrophic debris-flow, alluvial, or slushflow deposits, albeit derived partly from upslope till. Such episodic high-energy redeposition seems far more common in mountainous terrain than is generally acknowledged in the literature. Yet unless the texture and surface morphology of deposits have been made visible by deep roadcuts and clear-cut logging, postglacially redeposited till may be difficult to distinguish from primary *in-situ* till. We suggest below that high-energy postglacial slope processes such as debris flow are responsible for many till-like diamicts and vaguely morainelike topography in many mountainous areas of New England. Such deposits may be mistaken as evidence that local glaciers advanced from valley heads after the continental icesheet had disappeared.

GREEN MOUNTAINS, VERMONT

The Mount Mansfield ridge and areas immediately to the east and west in the Green Mountains (fig. 3) consist of mica schist and phyllite (Albee, 1957; Christman, 1959; Christman and Secor, 1961; Doll and others, 1961). Foliation strikes and fold axes trend generally north-northeast. Drift stones of granite and volcanic rock are erratic to this bedrock and could only have been imported by overriding continental ice. Striae and perched boulders show that the summit ridge of Mount Mansfield (alt 1340 m)—the highest ridge of the Green Mountains—and other high ridges in Vermont were overridden by continental ice during the last glaciation (Christman, 1959; Stewart and MacClintock, 1969). Drift containing scarcely weathered, striated stones lies at all altitudes but mostly in lowlands.

Stewart (1961), Connally (1968), and Stewart and MacClintock (1969) saw no moraines or other evidence for post-icesheet local glaciation and interpreted all drift as a product of glaciation and deglaciation

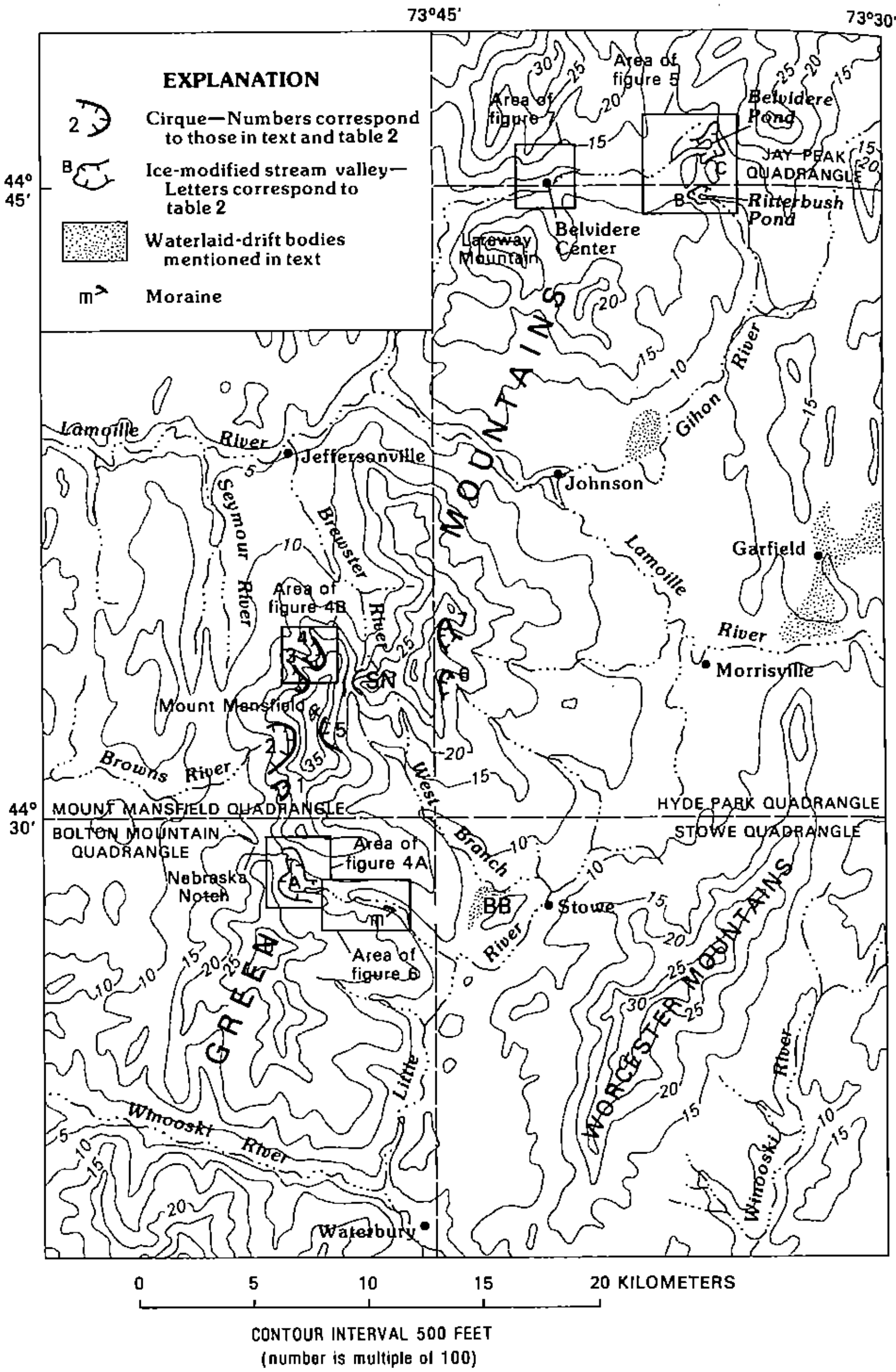

Fig. 3. Map of Mount Mansfield area, Green Mountains, Vermont, showing quadrangle boundaries (dashed lines), some glacial features, and locations discussed in text. BB, Barrows Brook; SN, Smugglers Notch. Generalized contours redrawn from U.S. Geological Survey map sheet Lake Champlain, Vt. and N.Y. 1:250,000 quadrangle.

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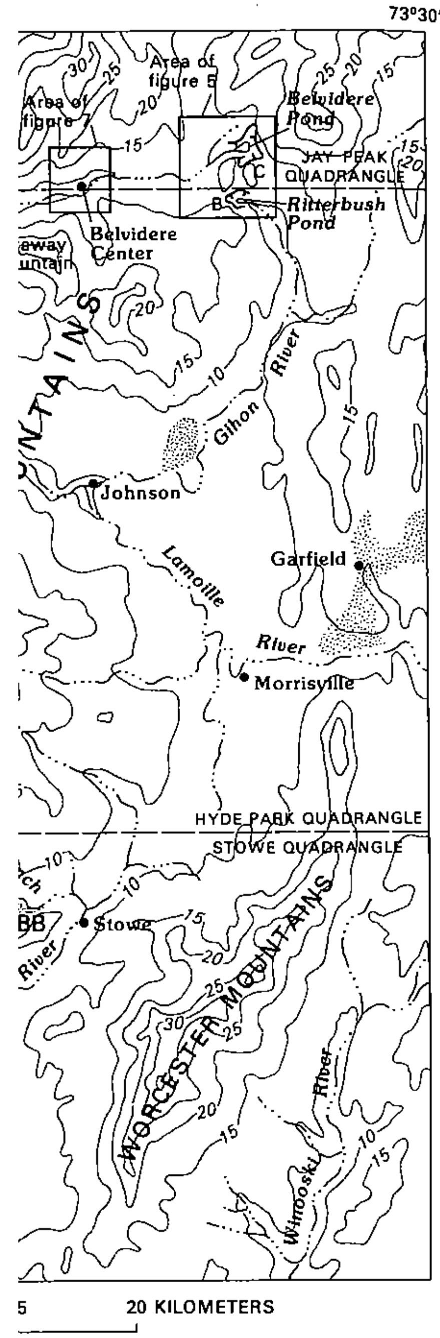

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Vt. and N.Y. 1:250,000 quadrangle.

by Laurentide ice. But Wagner (1970) inferred that local glaciers formed at low altitude in the Green Mountains after the icesheet melted. The case for post-icesheet local glaciation in Vermont (Wagner, 1970, 1971; Connally, 1971) rests on three arguments concerning valleys near Mount Mansfield (fig. 3): (1) the heads of some tributary valleys resemble cirques, and lakes within them resemble tarns; (2) these valleys contain moraines resembling those typically built by local mountain glaciers; (3) some deltas downvalley from these moraines seem too large to have been derived from a retreating icesheet. Some have regarded this evidence as compelling (Connally, 1971; Borns and Calkin, 1977; Davis and Davis, 1980; Bradley, 1981). Our field reexamination shows instead that this field evidence is ambiguous and that the arguments depend on dubious analogy. Indeed the Green Mountains give little or no evidence that local glaciers formed after continental ice disappeared.

Valley heads.—Wagner (1970) called the heads of Nebraska, Ritterbush, and Belvidere valleys "cirques" (fig. 3), yet these basins differ markedly from the indisputable cirques of the Presidential Range, Mount Katahdin, and numerous mountain ranges of western North America. Ideally a cirque is an inverse quarter-sphere having steep, upward-concave sides and headwall that lead from a much less-steep upland and which overlook a distinct cirque floor. Strongly contrasting rock belts may influence the shape of cirques, but in the Green Mountains and other areas discussed, the valley heads are in relatively homogenous rock or are cut across structural trends. The cirque form can be complex where several adjacent glaciers carve a compound cirque, and low-altitude cirques may be far less developed than higher altitude cirques nearby. But even in compound or poorly developed cirques the nondigitate bowl shape is evident, spurs and upland surfaces are distinctly truncated, any minor tributaries hang. In contrast, the so-called "cirques" at the heads of Nebraska, Ritterbush, and Belvidere valleys range from gently to steeply sloping; side profiles are diversely straight, concave up, convex up, or irregular; these forms blend into one another as well as into upland forms; tributaries are digitate and generally not hanging, and intervening spurs are not trucated (figs. 4A and 5). Numerous slopes in the area, such as Smugglers Notch and Laraway Mountain, were beveled even steeper by the through-flowing icesheet. The heads of Nebraska, Ritterbush, and Belvidere valleys (alt 315-400 m) are 100 to 300 m lower than many nearby valley heads that are not at all cirqueform—such as the heads of the West Branch Little River and Brewster Creek valley (near Nebraska valley), Lockwood and Calavale Brooks (near Belvidere and Ritterbush Ponds), and the south fork of Wild Brook (near Ritterbush Pond).

Evident cirques do lie in the Green Mountains above alt 670 m. The head of an unnamed tributary of Seymour River on the northwest flank of Mount Mansfield has conspicuously arcuate, quarter-bowl-shaped sides and headwall above alt 670 m [2200 ft] (fig. 4B); below that

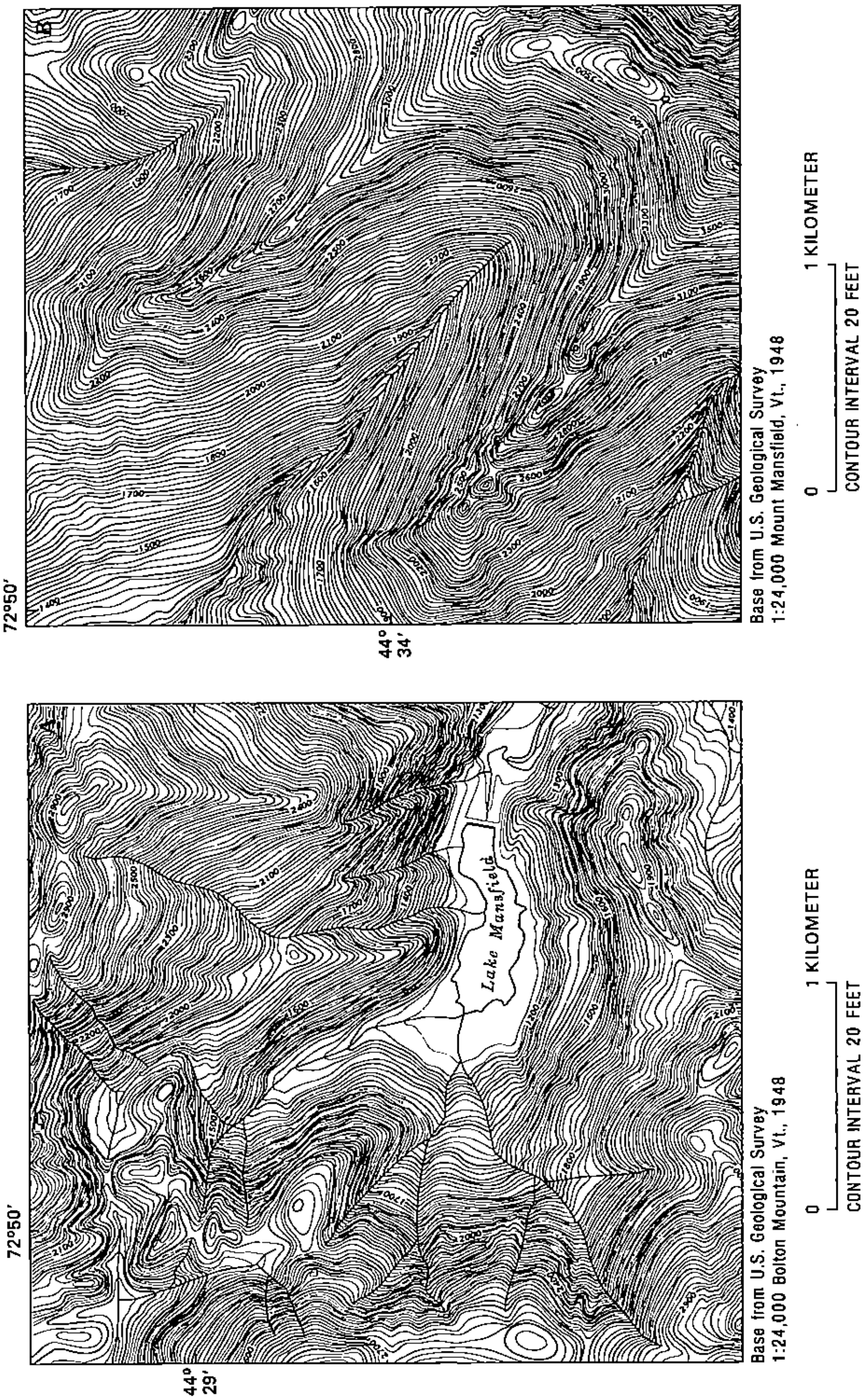

Fig. 4. Tributary heads on Mount Mansfield. (A) Low-altitude noncirque head of Nebraska valley. (B) Cirque basins of upper flanks of Mount Mansfield.

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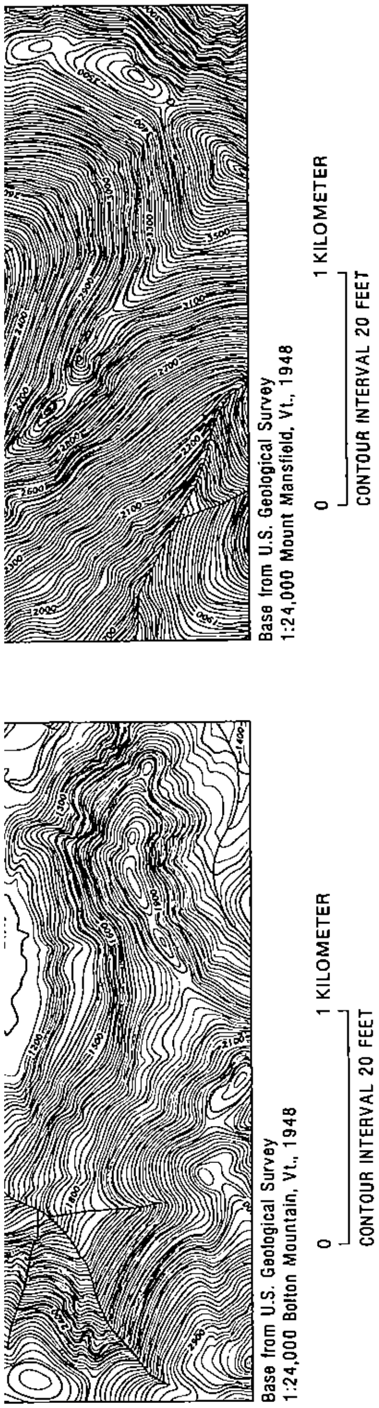

Fig. 4. Tributary heads on Mount Mansfield. (A) Low-altitude noncirque head of Nebraska valley. (B) Cirque basins of upper flanks of Mount Mansfield.

altitude the straight, V-shaped valley sides are a classic stream-eroded topography. A broad basin on the east flank of Mount Mansfield (containing head of ski area) is only crudely cirqueform, but the arcuate, steep, concave-up northeast-facing wall is cirque shaped above alt 900 m and overlooks a narrow floor at 850 to 900 m. Other weakly formed but distinctly arcuate cirques are cut in the west flank of Mount Mansfield (Stevensville Brook, a headwater of Brown's River), and other small cirque basins on the west side of the mountain lie above alt 730 m [2400 ft].

Table 2 summarizes numerical data in the Green Mountains for definite cirques, possible cirques, and low-altitude pseudocirque valley heads; it compares these data to those from undisputed cirques in New Hampshire and Maine. The definite cirques are simpler in form, average about 500 m higher in altitude, and have maximal slopes about 10° steeper than the low-altitude pseudocirques. The values for the definite cirques compare well with those from definite cirques in New Hampshire and Maine; the low-altitude basin data do not. Two basin heads (table 2, nos. 6 and 7) on the east side of the Mount Mansfield ridge are more cirquelike than Wagner's low-altitude basins but are lower by about 250 m average than the definite cirques. These basins may be rather poorly developed cirques.

The northwest-facing definite cirques on Mount Mansfield are cirqueform only above 670 m; that facing southwest only above 760 m. The snowline that controlled cirque glaciers therefore must have been at or above 670 m [2200 ft] in north-facing valley heads and at or above 760 m [2500 ft] in south-facing valley heads. Thus in this region cirques would be difficult to explain in valley heads at alt 315 to 400 m.

Mountain passes above the Nebraska, Ritterbush, and Belvidere basins were overwhelmed and buried by some 1500 m or more of Laurentide ice that streamed through the basin heads during the glacial maximum (Stewart and MacClintock, 1969, p. 41) perhaps for millennia. Belvidere valley head in particular has the form of an icesheet-reamed glacial trough. During the last and probably earlier glaciations the icesheet crossing the Green Mountain ridge had ample opportunity to widen these and other preglacial stream-valley heads out into their broad, crudely U-shaped, steep-headed, steep-sided forms.

Drift and moraines.—Some features in the Green Mountains have been called alpine-glacier deposits, yet few of these landforms closely resemble those built at the margins of valley or cirque glaciers in the western United States (for example, Porter, Pierce, and Hamilton, 1983). Some of the features that Wagner (1970) initially thought were moraines built by local glaciers are not distinctly arcuate ridges but are vague mounds similar to icesheet ground moraine all over New England. We agree with Wagner (1971) and Connally (1971) that these deposits probably are from the icesheet.

The largest and most convincing of Wagner's (1970, 1971) inferred alpine moraines is on the south side of Nebraska valley 4.5 to 5

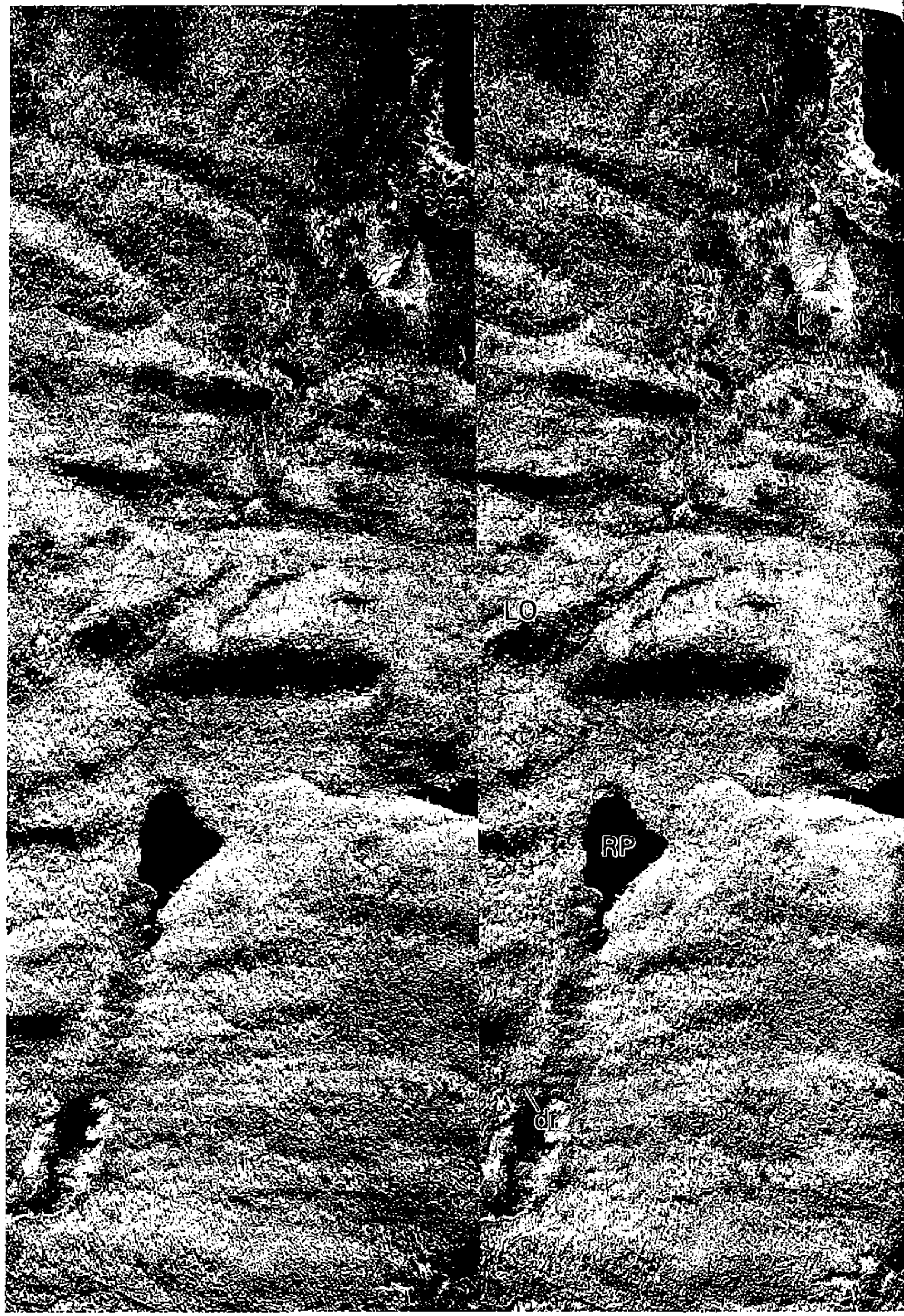

Fig. 5. Stereogram of three vertical photographs showing heads of Ritterbush and Belvidere valleys, Vermont. BP, Belvidere Pond; RP, Ritterbush Pond, LO, lake outlet

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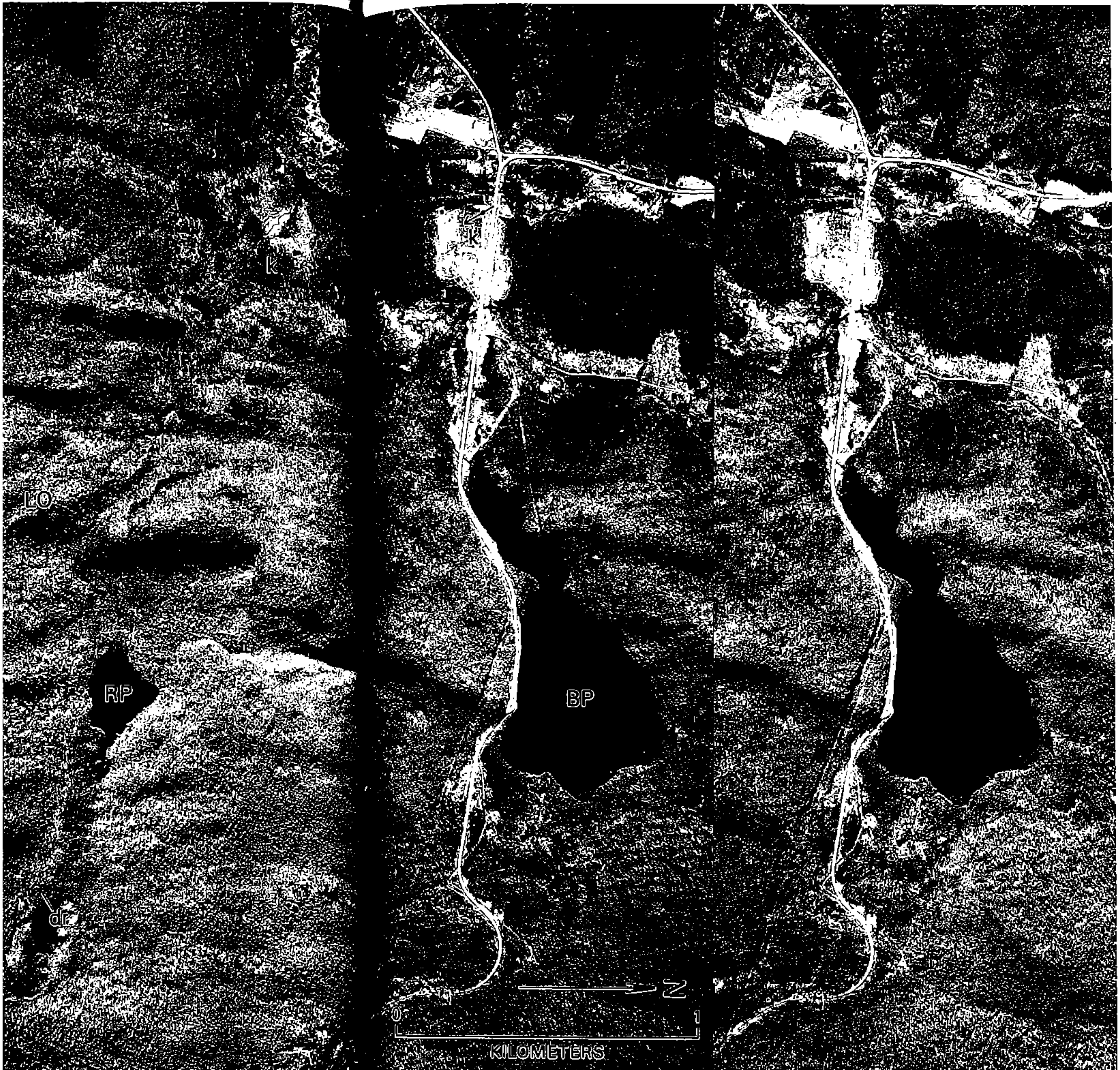

ɔhotographs showing heads of Ritterbush and
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from Belvidere to Ritterbush valley; dr, drift ridges; k, kames. Photographs by Aero-Graphics, Bohemia, N.Y., series VT7420, nos. 15-017 to 15-019 (1974).

TABLE 2

Statistics for cirques and low-altitude pseudocirque basins near Mount Mansfield, Vermont compared to cirques in New Hampshire and Maine

Character	Green Mountains, Vermont													Presidential Cirque Average**	Katahdin Cirque Aver-age***	West-central Maine Cirque Average†
	Definite Cirque*						Possible Cirque*			Low Basin Head*						
	1	2	3	4	5	Ave	6	7	Ave	A	B	C	Ave			
Azimuth (degrees)	230	270	315	315	065		145	160		105	115	255				
Maximum altitude of cirque rim (m)	1160	1190	1160	1130	1160	1160	825	975	900	795	425	490	570			
Minimum altitude of cirque floor (m)	670	640	520	520	700	610	640	655	650	345	315	350	335	>770††	1150	
Maximum height rim to floor (m)	490	550	640	610	460	550	215	320	270	450	130	140	240	235	350	
Schrund altitude (m)	1005	945	795	795	975	905	700	670	685	360	335	365	355	1245	995	1000
Steepest long slope rim to floor (degrees)	33	28	30	27	40	32	18	21	20	22	18	20	21	37	44	
Shape in plan†††	s	s	so	s	so		s	s		d	sd	sd				

*Identified on figure 3
**Goldthwait (1970a, table 1)
***Davis (1976, table 3)
†Borns and Calkin (1977, table 2)
††Goldthwait gives average rather than maximum height
†††s, simple open-ended bowl; so, simple one-sided bowl; d, digitate; sd, somewhat digitate

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[illegible]	[illegible]	[illegible]	520	520	700	610	640	655	650	345	315	350	335	>770††	1150	
Maximum height rim to floor (m)	490	550	640	610	460	550	215	320	270	450	130	140	240	235	350	
Schrund altitude (m)	1005	945	795	795	975	905	700	670	685	360	335	365	355	1245	995	1000
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Shape in plan†††	s	s	so	s	so		s	s		d	sd	sd				

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km from the valley head (fig. 5). A sharp-crested end moraine 5 to 15 m high extends less than 0.5 km to the valley side; it is perched atop a body of waterlaid gravel and sand 20 to 30 m thick that generally becomes coarser upward. The moraine consists of diamict with boulders as large as 1.2 m but merges downvalley into bedded, gently foreset-bedded, sorted sand and gravel—evidently outwash and deltaic facies. The ridge is similar in form to many small last-glacial alpine moraines in western United States; Connally (1971) and several others including us have inspected this ridge and agree that it indeed has the general form of an alpine-glacial moraine. Yet arcuate moraines can accumulate along any tongue of ice confined by valley walls regardless of whether the tongue issues from a cirque glacier, a local icecap, or an invasionary icesheet. Arcuate moraines similar to and even larger than that in Nebraska valley were built by valley glaciers fed by a dwindling icecap in northern Newfoundland (Waitt, 1981, fig. 3), by receding Cordilleran ice in northern Washington (Waitt, ms, p. 127–129; Waitt and Thorson, 1983, fig. 3–4), and by other icesheet-fed tongues such as in Greenland (Ten Brink and Weidick, 1974). Gravel clasts in the Nebraska valley moraine and in associated waterlaid drift within 1 km both upvalley and downvalley are mostly locally derived schist and micaceous gneiss, but 10 to 20 percent are erratic igneous and sedimentary clasts. The mixed lithology of gravel clasts and the form of the ridge equally well admit of two explanations: (1) that the moriane and associated drift were built by an icesheet tongue fed from over the valley head; or (2) that a local glacier re-formed after the icesheet and removed mixed-lithology icesheet drift from the valley head to the moraine. South-southeast-trending striations on smoothed bedrock 1.5 to 2 km downvalley of the moraine (fig. 6) show in any case that when the icesheet was more extensive its basal flow had been directly down Nebraska valley. The Nebraska valley moraine is scarcely more remarkable than icesheet deposits scattered about the region. Although the moraine may be explained as a hypothetical mountain glacier, it can be wholly explained by the waning icesheet that unarguably occupied the valley and whole region.

The northern and most cirquelike arm of Nebraska valley hangs 250 m above the head of the main trough (Lake Mansfield) (fig. 4A). This high valley floor is littered mainly by angular to subangular locally derived mica schist and micaceous gneiss, but it also contains subangular to rounded clasts of imported granite, metavolcanic rocks, and mica-free gneiss. Some of the surrounding valley walls are mantled by drift, so some erratics may have arrived at the valley floor in postglacial debris flows. But the rounded erratics are at least permissive evidence that the last ice to waste even from the highest part of Nebraska valley was an arm of the icesheet.

The "moraines" near the head of Ritterbush valley (fig. 5) are discontinuous mounds or short ridges that are not moored to the valley sides by lateral moraines; they are not continuous convex-downvalley

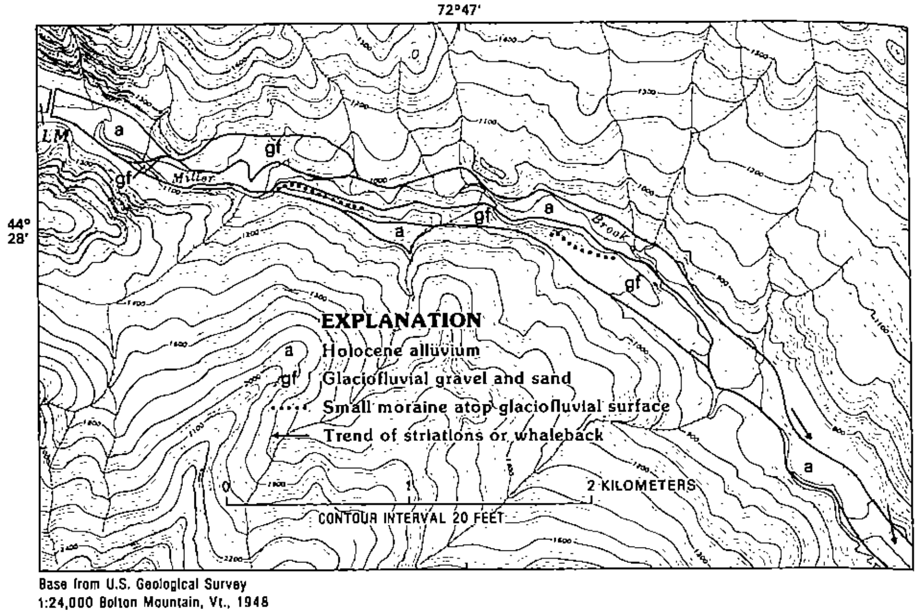

Fig. 6. Drift below Lake Mansfield near head of Nebraska valley.

arcs suggested by Wagner's (1970, fig. 1) schematic map symbol. Ritterbush Pond and smaller depressions just downvalley are surrounded by ice-scoured phyllite and schist ridges that trend north-northeast, parallel to the typical strike of bedrock foliation in that area and to foliation in adjacent areas mapped by Albee (1957). Downvalley of the pond are only three parallel ridges, each less than 75 m long and less than 5 m high. These trifling deposits were probably formed by an upvalley-receding distributary glacier fed by the icesheet from over the valley divide. Wagner infers another moraine 3.5 km from the head of Ritterbush valley; but this feature is part of an irregular ice-contact body consisting of mixed-lithology drift like that of scores of similar bodies at both higher and lower altitudes in the broad vale between the Green Mountain and the Worcester Mountains, well beyond any possible influence by mountain glaciers.

In Belvidere valley Wagner inferred two "moraines," one damming Belvidere Pond and the other 6 km downvalley at Belvidere Center. But they seem not to be moraines at all. The angular phyllite boulders at Belvidere Center and hummocky topography extending up the north valley side to an arcuate scar are debris of a postglacial landslide (fig. 7). The material damming Belvidere Pond is nearly flat, postglacial valley-floor alluvium (fig. 5); rising above this level are several hummocky or kettled deposits, evidently ice-contact outwash or

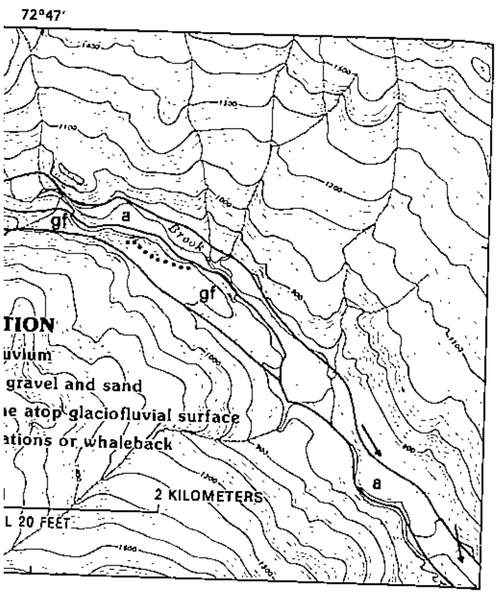

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Fig. 7. Stereogram of two vertical photographs showing landslide deposit at Belvidere Center, Vt. Slide area outlined; arrow shows direction of movement. Photographs by AeroGraphics, Bohemia, N.Y., series VT7420, nos. 14-103 & 14-104 (1974).

subaqueous-fan material. This moderately sorted sandy gravel contains diverse metamorphic and sedimentary rocks, whereas the valley head is carved out of phyllite and schist. Drift mounds, kettled kames, and apparently glaciofluvial deltas related to the withdrawal of the icesheet lie scattered about the valley between Belvidere Pond and Belvidere Center. Most or all of these deposits contain diverse gneissic stones that are erratic to the local schist and phyllite bedrock. About 1 km downvalley from Belvidere Pond, for instance, a distinct ridge of highly mixed-lithology, waterlaid, kettled drift has foreset bedding dipping southeastward—nearly upvalley and directed toward the outlet of a temporary ice-dammed lake (LO on fig. 5). The last glacier to recede from the valley seems to have been a wasting icesheet tongue fed from the west, not local ice flowing from the valley head on the east.

Not even the distinct cirques that head above alt 700 m on the west flank of Mount Mansfield (fig. 3; table 2) are floored with alpine drift. At one of these basin heads drained by the headwaters of Browns River (fig. 3, cirque 2), several discontinuous ridges descend from alt 750 to 300 m and trend downslope parallel to the deeply incised brook. These ridges consist of nonsorted drift that contains boulders as large as 3 m and is nearly identical to till that thickly blankets the lower reaches of Browns River and other valleys to the west and northwest. About three-fourths of the gravel-sized clasts are of schist similar to local bedrock; the rest are imported clasts of basalt as well as rounded, far-traveled quartzite and gneiss, some of them faceted and striated.

The drift ridges in Brown's River headwaters are not arcuate and astride the valley but lie closely parallel to each other and to the narrow incised inner valley. The distribution and shape of the ridges indicate that alpine glaciers were not the agents that redistributed the icesheet drift downslope. Two inferred processes explain the valley-floor drift: (1) much of the drift and depositional topography was formed by the icesheet, which deposited its drift broadly across lowlands to the west and north; (2) the ridges that closely parallel streams are levees built by debris flows just after deglaciation, when loose icesheet drift was unchecked by vegetation and susceptible to downslope redistribution. Morphologically these deposits resemble the storm-generated leveed debris-flow deposits such as mapped elsewhere by Beaty (1963) and by Wells and Harvey (1987, figs. 2, 3). If such debris flows occurred soon after deglaciation before the brooks became incised 30 to 60 m in the present inner valleys, the flows need only have been 1 to 10 m deep.

Waterlaid drift.—Wagner (1970) correlates an ice-contact body in Ritterbush valley with a delta 8 to 9 km downstream along the Gihon River (fig. 3) and takes this as evidence for a significant alpine-glacial episode after icesheet glaciation. But both deposits can be logically attributed to icesheet tongues receding up the Gihon and Ritterbush valleys. Wagner's inference that many drift bodies are "anomalously" large does not consider that sub-icesheet drainage, unvegetated drift-mantled slopes, and loose drift bodies on the valley floor were abundant,

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ready sources of sediment for transport by water associated with the wasting icesheet—such as has recently been shown for Malaspina Glacier, Alaska (Gustavson and Boothroyd, 1987).

Masses of such glaciofluvial sediment abound in valleys on both sides of the Green Mountains. These valleys contain hundreds of scattered deposits of waterlaid gravel, sand, and silt that overlie bedrock and till at various heights as much as 250 m above valley floors (Connally, 1968). Some of the waterlaid deposits such as at Barrows Brook (fig. 3) and that cited above merging with the Nebraska valley moraine are close to mountain ridges, but most of them such as at Garfield (fig. 3) are far from any conceivable influence from local glaciers. Many of these deposits contain hummocky or kettled areas interspersed with flat-surfaced terraces or deltas; adjacent to some of these deposits are valley-floor swamps closely resembling "dead-ice sinks" that Fleisher (1986) attributes to icesheet stagnation in high-relief terrain. These deposits must be related to the receding icesheet; significant accumulation seems impossible at most of these sites except while lingering ice or ice-dammed lakes broadly occupied the Winooski and Lamoille River drainages.

Conclusion.—If any post-icesheet mountain glaciation occurred in the northern Green Mountains, it left no diagnostic features. All field evidence in mountain valleys is readily accounted for by icesheet deglaciation and by later mass wasting and alluviation.

MOUNT KATAHDIN, MAINE

Far-traveled north-derived erratics lying within 45 m altitude of the summit show that Mount Katahdin, the highest mountain in Maine (alt 1605 m), was overridden by a continental icesheet (Tarr, 1900; Antevs, 1932; Davis, 1976, p. 69). The freshness of erratics and polished bedrock surfaces on Mount Katahdin, the thinness of soils there, and theoretical ice profiles extrapolated back from recessional ice positions on the Maine coast indicate that the maximal late Wisconsin icesheet buried Mount Katahdin (Davis, 1976, p. 69–75; Davis, 1978, 1983, 1988). Quantitative reconstruction of the icesheet surface from the characteristics of eskers (Shreve, 1985) suggests that later after considerable deglaciation the upper half of Mount Katahdin had emerged as a nunatak.

Great cirques carved into the flanks of Mount Katahdin have long been considered a product of post-icesheet local glaciation (fig. 8). The east and north flanks of Mount Katahdin are cut by 5 compound cirques. The steepest long slopes of cirques range 35° to 60°; by these and other numerical data (Davis, 1976, table 3) the Katahdin cirques compare well with cirques from the Presidential Range (Goldthwait, 1970a, table 1) and with cirques inferred here from the Green Mountains (table 2). Virtually all previous researchers have thought that the cirques and separating arêtes on Mount Katahdin's east side (fig. 9) are too fresh for them not to have been sharpened by local glaciers after the supposedly smoothing effects of the overriding icesheet (Tarr, 1900; Antevs, 1932; Thompson, 1960–61; Caldwell, 1966, 1972, 1980). As thus recently proclaimed: "the [arêtes] could not, in their present form, have withstood the effects of an overriding icesheet" (Caldwell, Hanson, and Thompson, 1985, p. 55). Yet sharpness of cirques is not diagnostic of post-icesheet cirque glaciation; for as noted above, ragged arêtes and well-formed cirques in the North Cascade Range of Washington were scarcely altered by burial beneath hundreds of meters of the late Wisconsin Cordilleran icesheet (fig. 2). Various erosional and depositional evidence shows that the last ice to occupy the cirques of Mount Katahdin, as in the North Cascades, was an invasionary icesheet. Below is a summary of the Katahdin evidence from Davis (ms).

Roches Moutonnées.—At least three roches moutonnées embellish the floor of Northwest Basin (fig. 8) about 1.5 km from the cirque headwall. The long axes of these forms and glacial grooves etched into them trend south-southeast, slightly oblique to the trend of Northwest Basin cirque. The grooved, chattermarked stoss ends face downvalley; the ragged lee ends face the cirque headwall. These relations suggest that the last flow of ice was *up* the cirque by the regionally southeast-flowing icesheet and that no sizable cirque glacier reformed in Northwest Basin after the icesheet receded.

Cirque-floor drift.—Mount Katahdin is formed of the Katahdin Granite (biotite granite to granophyre), which is intruded into lower Paleozoic sedimentary and volcanic rocks that surround the mountain mass (Osberg, Hussey, and Boone, 1985). Thus all sedimentary and volcanic stones in drift on Mount Katahdin are erratic. Gravel-size stones in the thin till sheet on the floor of Northwest Basin are dominantly granite, but 22 to 38 percent of them are sedimentary and volcanic erratics (table 3). Many of these stones are derived from the Paleozoic Seboomook and Matagamon Formations, bedrock units north of Mount Katahdin. Many of the granite clasts also may have been derived from the lower mountain flank northwest of the cirque, for Katahdin Granite bedrock extends 10 km north of Northwest Basin (Osberg, Hussey, and Boone, 1985). A post-icesheet cirque glacier such as envisioned by Antevs (1932) should have redistributed the imported drift back downvalley or overlaid it with erratic-free drift; if a cirque glacier hypothetically removed thick icesheet drift from the cirque floor it should have built conspicuous moraines just beyond the cirque. The presence of erratics in the thin cirque-floor drift and the absence of any notable moraine farther downvalley indicate that the last glacier to occupy Northwest Basin was the icesheet flowing up the cirque—the same southeastward ice-flow direction inferred from roches moutonnées in the basin.

Erratics are much less abundant in the thick drift on the floors of the four east- and north-facing cirques than in Northwest Basin cirque, yet some pebble-composition counts (North Basin, Great Basin, South Basin) show as much as 20 percent sedimentary and volcanic erratics

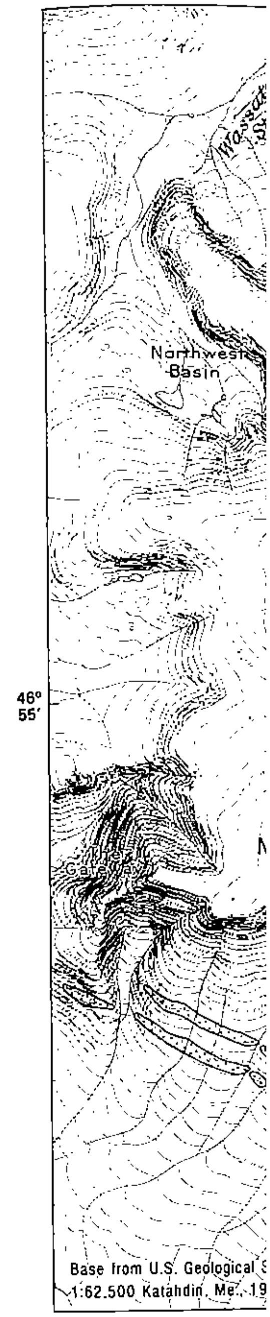

Fig. 8. Map of N
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Katahdin, Maine (1949

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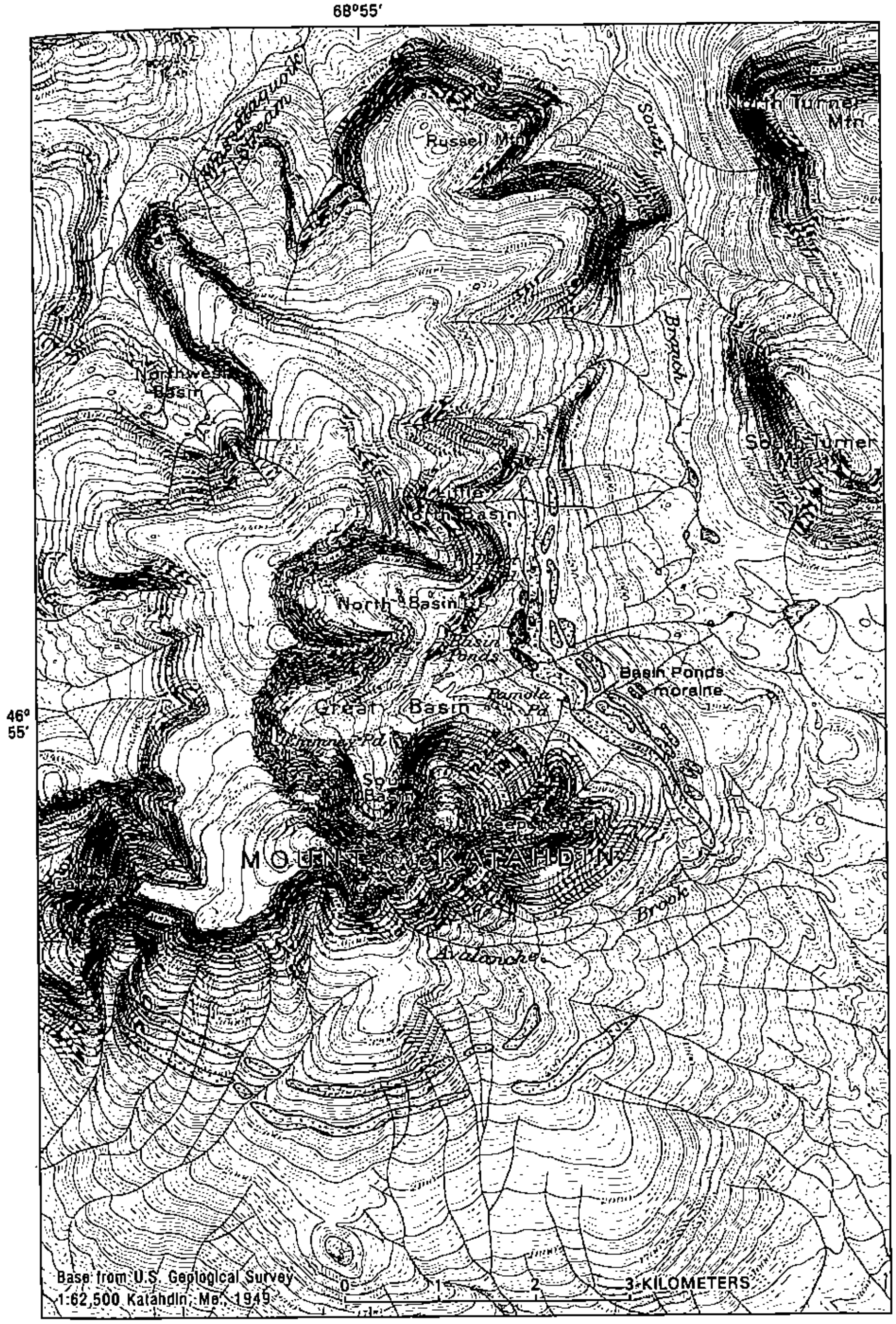

Fig. 8. Map of Mount Katahdin area, Maine, showing distribution of moraines (stippled areas) on east and south sides. Base from U.S. Geological Survey 1:62,500 Katahdin, Maine (1949) quadrangle.

Fig. 9. Aerial view of east side of Mount Katahdin, Maine, showing eastside cirques and Basin Ponds moraine and lower moraines (dashed lines).

Erratic s

Sample Number
K85
K86
K87
K88
K11
K42
K47
K56
K43
K44
K10
K32
K33
K34
K35
K55
K31
K54
K57
K58
K59
K60
K61
K62
K63
K64
K65
K66
K67
K68
K69
K70
K91
K92A
K92B
K93
K94
K71
K72
K81
K82
K83
K96
K97
K98
K99

*Summa fig. 13).
**Include chert, slate, mic

Fig. 9. Aerial view of east side of Mount Katahdin, Maine, showing eastside cirques and Basin Ponds moraine and lower moraines (dashed lines).

TABLE 3

*Erratic stones counted in pebble fraction of till, Mount Katahdin**

Sample Number	Location	Number Counted	Percent Erratic**
K85	Northwest Basin	100	38
K86		50	30
K87		50	22
K88		50	28
K11	North Basin	100	15
K42		50	20
K47		100	13
K56		100	9
K43	(Blueberry Knoll)	50	16
K44		50	14
K10	Great Basin	200	2
K32		200	3
K33		200	4
K34		100	3
K35		200	2
K55		200	2
K31	South Basin	100	4
K54		200	5
K57	Basin Ponds moraine, north	50	20
K58		50	24
K59		50	18
K60		50	14
K61		50	10
K62		50	32
K63		50	14
K64		50	26
K65		50	32
K66		50	16
K67		50	22
K68		50	24
K69		50	40
K70		50	32
K91	Basin Ponds, moraine, south	50	44
K92A		50	34
K92B		25	28
K93		50	24
K94		50	26
K71	Moraines east of B. P. moraine	50	42
K72		50	12
K81		25	48
K82		50	64
K83		50	68
K96	Moraines south slope Katahdin, east	50	42
K97		50	48
K98	Moraines south slope Katahdin, west	50	32
K99		50	24

*Summarized from detailed counts in Davis (1976, app. B; specific locations on his fig. 13).

**Includes 31 distinguished varieties of sandstone, siltstone, shale, meta-sandstone, chert, slate, mica schist, chlorite schist, and volcanic rocks.

(table 3). Erratics among the cobble-boulder fraction are also easily distinguished in trailcuts in these cirques. Thompson (1961, p. 470) and Caldwell (1966, p. 53) interpret the lower abundance of erratics on the floors of the east-facing cirques to mean that cirque glaciers removed material after the continental ice disappeared. An alternative explanation is basal freezing and regelation by which till is incorporated on the upglacier side of bedrock obstructions and deposited on the downglacier side (Boulton, 1972, p. 5; 1974, p. 55). Thick continental ice flowing over Mount Katahdin, an obstruction to its flow (high-pressure area in glacier bed), incorporated mainly granite as it flowed up the northwest mountain flank, and then it deposited mainly granite clasts in the east-facing cirques that constituted hollows (low-pressure areas) in the glacier bed. It is also unlikely that this drift on the floors of east-facing cirques was derived from postglacial mass wasting of drift deposited on cirque walls. The cirque walls are too steep to have accumulated much drift, and later mass wasting from the steep rock slopes would inevitably consist mostly of angular local-rock debris, as some inferred mass-wastage deposits near valley sides indeed do. The smaller abundance of erratics in the east-facing cirques is no proof of mountain glaciers after the icesheet disappeared (Davis, ms).

Moraines.—The distinct Basin Ponds moraine (Hamlin, 1881; Tarr, 1900; Antevs, 1932; Thompson and Borns, 1985) extends for 5 km approximately on contour at 740 m (2400 ft) altitude across the eastern slope of Mount Katahdin as a continuous hummocky ridge that dams Basin, Depot, and Pamola Ponds (figs. 8, 9). The moraine consists of mixed-lithology till but also contains many rounded granitic boulders as much as 6 m in diameter. The Basin Ponds moraine has been interpreted as a terminal moraine deposited by late Wisconsin mountain glaciers flowing eastward from Mount Katahdin (Tarr, 1900), as a medial moraine between such mountain glaciers and continental ice to the east (Antevs, 1932; Caldwell, 1966, 1972), or as a lateral moraine marking the maximum late Wisonsin continental ice limit against an ice-free Mount Katahdin (D. Grant, oral commun., 1976; Caldwell and Hanson, 1986). But the morphology of the moraine and the erratic component of its stones do not support the first two ideas, whereas the freshness of weathering and soils on Mount Katahdin dispute the third (Davis, 1978, 1983).

Basin Ponds moraine does contain many large granitic boulders, but they may have been transported by an icesheet flowing around the mountain as readily as by local glaciers from the eastside cirques. In any case, the pebble fraction is 10 to 44 percent erratic (table 3). The moraine neither is convex eastward nor does it descend eastward like a local mountain-glacier moraine; instead it is convex westward and roughly follows a contour along the east slope of Mount Katahdin. There is little space between it and Keep Ridge for a hypothetical former cirque glacier or even a large drainage channel (fig. 8). The

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moraine extends both north and south *beyond* the mouths of the three east-facing cirques. The Basin Ponds moraine must therefore have been built by receding late Wisconsin continental ice that occupied the broad valley to the east when Mount Katahdin had emerged as a nunatak during the esker-forming stage (Shreve, 1985). Thus both the lithology and form of the Basin Ponds moraine are wrong for its deposition by hypothetical cirque glaciers flowing eastward from the mountain.

Several smaller moraines lie on the east mountain flank just downslope from the Basin Ponds moraine. Some of these ridges are vaguely arcuate, but generally they are parallel to each other and to the Basin Ponds moraine. They contain 42 to 68 percent erratic pebbles (table 3), even more than the Basin Ponds moraine, and many of their stones are striated and faceted. These moraines must have been built by the west margin of the continental glacier as it receded downslope and eastward from the Basin Ponds moraine. Similar moraines also lie downslope of the Basin Ponds moraine below Keep Ridge and below Little North Basin. These moraines are significant, because they lie both south and north of the three east-facing cirques and could not possibly have been built by cirque glaciers. Thus for several reasons we reject Thompson's (1961) speculation that moraines outside the Basin Ponds moraine were formed by cirque glaciers that overrode the Basin Ponds moraine.

Several closely spaced moraines lie along the noncirqued south slope of Mount Katahdin (fig. 8) (Davis, ms and, 1983; Davis and Davis, 1980). Nearly continuous ridges extend for about 8 km and descend in altitude from 730 m [2400 ft] in the east to 570 m [1850 ft] in the west. Many erratics, some with striated and faceted faces, lie on the moraines. The high percentages of erratic pebbles (table 3), including red shale not found higher on the mountain, indicate that the moraines were built by a tongue of continental ice flowing around Mount Katahdin. These erratics and the absence of cirques on the south side of the mountain make deposition of these moraines by local glaciers impossible.

Near the mouth of Great Basin and South Basin, a hummocky landform named "Bears Den moraine" by Tarr (1900) encloses Dry Pond. Tarr interpreted this feature as a late-Wisconsin end moraine deposited by a cirque glacier from South Basin. The landform consists of a depression surrounded by mounds rather than continuous morainal ridges. The feature seems to have been cut by a postglacial stream, which concentrated large and rounded granitic boulders as a lag deposit by piping out fine particles. The feature offers no proof of cirque glaciation (Davis and Davis, 1980).

Thompson (1961) and Caldwell (1966) described a smooth and subdued mound east of Chimney Pond, interpreting it as an end moraine deposited by a cirque glacier in South Basin after disintegration of the continental icesheet. The mound is embellished by many rounded granitic boulders as large as 3.5 m in diameter; only 2 percent of the pebbles are erratic. Yet the mound neither has a ridgeform like Basin Ponds moraine nor is it arcuate: it could easily be a chance icesheet deposit and thus offers no proof of cirque glaciation.

Blueberry Knoll at the mouth of North Basin has been interpreted as an end moraine of a cirque glacier (Caldwell, 1966). The knoll is covered by till composed mainly of granitic boulders as large as 3 m in diameter, but the pebble fraction is as much as 16 percent erratic (table 3). Seismic profiles show an average thickness of 2 m of till over bedrock on Blueberry Knoll and a greater thickness on the adjacent cirque floor (P.T. Davis, unpub. data, 1976). Therefore Blueberry Knoll is not an end moraine but is a bedrock high capped by icesheet till.

No lateral and looped end moraines lie on cirque floors (Davis, ms, fig. 12); instead the cirques contain only formless till that near steep valley walls is overlapped by bouldery talus and protalus ramparts. A suggestion that ridgelike deposits in cirques have been subdued by erosion (Thompson, 1961, p. 471) is inconsistent with the excellent preservation of the Basin Ponds moraine and other moraines on Mount Katahdin (fig. 9). The speculation by Tarr (1900) that moraines lie hidden beneath dense forest does not hold for North Basin, whose floor is nearly unvegetated. Indeed, Davis's (1976) many traverses across all cirque floors with air photos in hand did not reveal morainal ridges in any cirque basin. Caldwell's (1972, fig. 1) map exaggerates selected drift ridges on cirque floors by ignoring hundreds of other hummocks within and outside the Mount Katahdin cirques.

Discussion and conclusion.—We see no reason why Northwest Basin should have a different glacial-deglacial history than the eastern cirques as suggested by Caldwell, Hanson, and Thompson (1985, p. 55). Taking into account its relatively small size appropriate to its somewhat lower altitude, we note that during unforested late Wisconsin time before the icesheet arrived Northwest Basin would have intercepted "blowover" snow from a large adjacent accumulation area on the west; moreover its northwest aspect favored a low rate of ablation for any glacier there.

We also see neither evidence nor reason for the deductive speculation by Caldwell, Hanson, and Thompson (1985, p. 55) that a marine calving bay to the east provided precipitation sufficient to reestablish cirque glaciers at Mount Katahdin. The marine limit is no closer to Mt. Katahdin than about 50 km and was confined to a fairly narrow and shallow arm. Calving-bay dynamics could conceivably have influenced readvances of lowland ice, but it is hard to see how any slight effects of increased precipitation and wind-drifted snow could have compensated for a snowline that must have lain hundreds of meters above cirque floors at Mount Katahdin during deglaciation of the icesheet during the esker-forming stage (Shreve, 1985).

We conclude that there is no definite evidence of post-icesheet mountain glaciation at Mount Katahdin. All the erosional features, drift, and moraines are most readily accounted for by the sequence of icesheet glaciation and deglaciation following the last significant mountain glaciation. These conclusions from Mount Katahdin accord with those of Borns and Calkin (1977) based on similar data from mountains in west-central Maine.

PRESIDENTIAL RANGE, NEW HAMPSHIRE

Till and scattered erratics at the summit of Mount Washington (alt 1918 m) and other summits of the Presidential Range (fig. 10) include granite and other igneous rock types (Goldthwait, 1939, 1976), where local bedrock is schist (Billings and others, 1979). These scarcely weathered erratics show that the entire range was overridden by the late Wisconsin continental icesheet. In cirques of the Presidential Range, uphill-facing roches moutonnées, drift of alien provenance, and an absence of looped end moraines have long been taken as evidence that local glaciers did not occupy the cirques after icesheet deglaciation (J.W. Goldthwait, 1913, 1916; R.P. Goldthwait, 1939, 1940). Goldthwait (1970a, table 2 and p. 90–91) argues convincingly that the nearly ubiquitous drift sheet on cirque floors contains gravel clasts that are 32 to 72 percent rock types imported from the north and thus that the most recent glacier in the cirques was the icesheet. Bradley (1981, 1982) challenged this idea on the basis of the lithology of material he considers to be till on the lower north flank of the Presidential Range and adjacent floor of Randolph Valley (fig. 10). He argued that because this deposit contains schist clasts derived from the mountains, local glaciers must have flowed northward to Randolph Valley from King Ravine and other cirques in the Presidential Range (fig. 10) after the icesheet had largely or entirely disappeared. Fowler (1984) reexamined the deposits in Randolph Valley and concluded that the deposits are colluvial and thus that Bradley's interpretations of post-Laurentide alpine glaciers are questionable. We generally agree with Fowler (1984), whose paper appeared after our 1983–84 fieldwork was complete; our independent observations complement Fowler's.

Cirque-floor drift.—King Ravine and Ravine of the Castles are distinct cirque basins on the north flank of the Presidential Range (fig. 10). Except for a conspicuous inactive rock glacier composed of large, angular schist boulders extending from the head of King Ravine, drift that mantles the broad floor of the ravine contains many rounded clasts of granite, diorite, and hornblende gneiss as large as 2 m. The bedrock sources of these crystalline rocks lie downslope or many kilometers to the north and northwest (Billings and others, 1979; Billings and Billings, 1975; Henderson and others, 1977). Ravine of the Castles is also floored by drift that contains at least 12 varieties of erratics (chiefly granite, diorite, gabbro, gneiss, and metavolcanic rocks) as pebble- to boulder-sized clasts, many being distinctly rounded, faceted, or striated. Similar material also lies on the floors of the noncirque valleys of upper Snyder Brook and Cascade Brook. Angular-debris fans in Ravine of the Castles and upper Cascade and Snyder Brooks contain micaceous gneiss or schist derived by mass wasting from steep valley sides; but the valleys are mainly floored by drift that contains abundant rounded erratics (chiefly

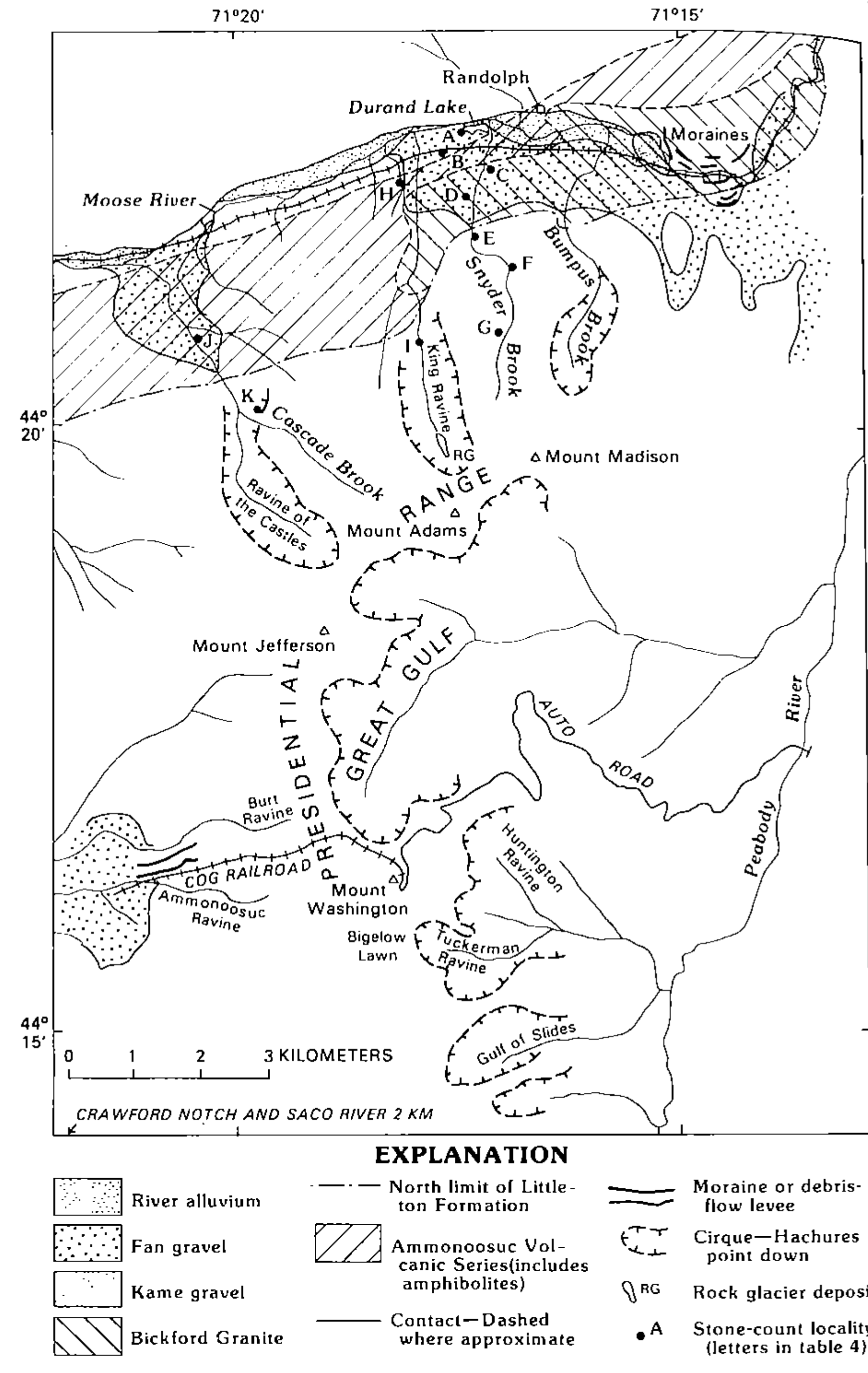

Fig. 10. Map of northern Presidential Range and adjacent Randolph Valley, N.H., showing generalized geology and distribution of cirques, debris fans, and other features.

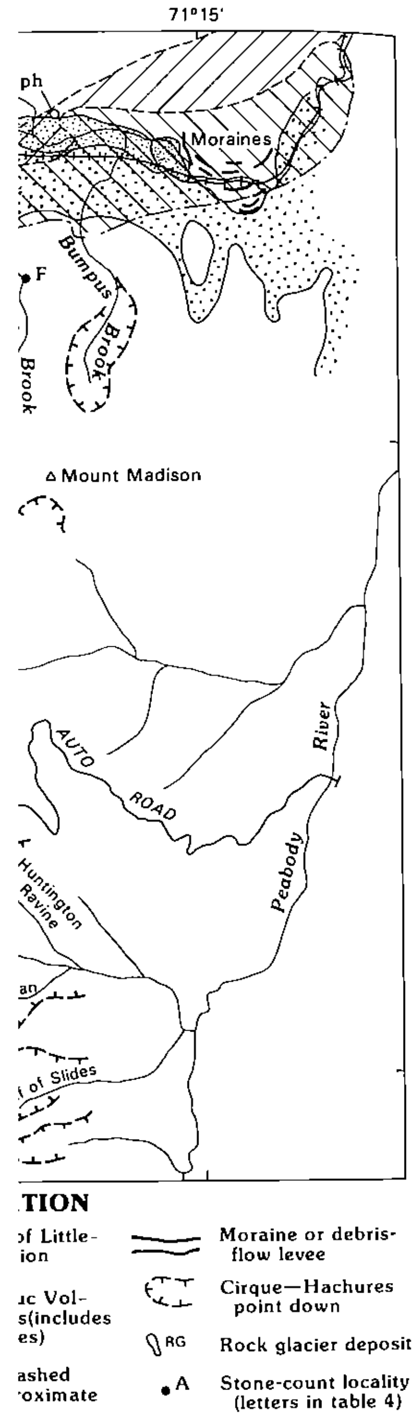

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TABLE 4

*Lithology of pebble fraction of till and fan material, northern Presidential Range**

Rock Types	Localities**										
	A	B	C	D	E	F	G	H	I	J	K
Muscovite-biotite schist***	19	17	12	8	8	4	4	24	8	24	—
Micaceous gneiss†	35	38	36	48	24	60	48	40	20	28	12
Pegmatites, quartzite, vein quartz, undiff. mafics††	37	31	28	32	44	28	28	32	40	48	40
Binary granite (Bickford)†††	—	—	—	—	—	—	—	—	—	—	—
Biotite-quartz monzonite and coarse pink granite (White Mountain Magma Series)[7]	4	3	—	4	4	—	4	4	8	—	12
Amphibolite and volcanics (Ammonoosuc Volcanic Series)[8]	3	9	20	4	16	8	12	—	20	—	32
Diorite (Lost Nation)[9]	1	1	4	4	—	—	—	—	—	—	4
Hornblende gneiss[9]	—	1	—	—	4	—	—	—	—	—	—
Quartzite[9]	1	—	—	—	—	—	4	—	4	—	—

*Expressed as percent of 25 pebbles counted (100 for localities A and B) with long axes 2 to 5 cm.

**Localities specified in figure 10: A, Durand Lake, south shore; B, railroad cut about 200 m south of Durand Lake; C,D,E,F,G, in ascending order, stream cuts in Snyder Brook valley; H, stream cut in fan below King Ravine; I, stream cut in lower part of King Ravine floor; J, stream cut in fan below Ravine of the Castles; K, stream cut in lower sidewall of Ravine of the Castles.

***Upper unit of Littleton Formation (Billings and others, 1979), possibly upvalley source for pebbles at all localities.

†Lower unit of Littleton Formation (Billings and others, 1979), possibly upvalley source for pebbles at all localities.

††Unknown source, but possibly derived from Littleton Formation.

†††Probable upvalley source for pebbles at localities A,B,C,D,H.

[7] Probable upvalley source for pebbles at all localities.

[8] Probable source for pebbles at localtiies E,F,G,I,K.

[9] Probable downvalley source at all localities.

granite, diorite, and metavolcanic rocks) as large as 2 m—identical to drift that mantles spurs between these valleys. The valley sides are much too steep to have accumulated much drift, and so it seems unlikely that the round-stone valley-floor drift could be postglacial debris-flow material from the valley sides. Stone counts in drift (Goldthwait, 1970a, table 2 and p. 90–91) show that the erratic content of drift on the floors of the Bumpus Basin, King Ravine, and Ravine of the Castles cirques ranges from 32 to 72 percent. Although our stone counts (Davis and Waitt, 1986) show somewhat lower erratic contents in the cirques (32 and 48 percent at loc. I and K, table 4), percentages of stones from definitely upvalley sources are also low (28 and 12 percent at the same localities). The texture and lithology of this drift also resembles that of the drift that broadly mantles the valleys flanking the Presidential Range on the east, north, and west. Similar erratic stones lie isolated or in drift at high altitude in the Presidential Range, as can be found above alt 1100 m along the Mount Washington Auto Road and on upland surfaces such as Bigelow Lawn.

The Goldthwaits' arguments that the most recent flow of ice in King Ravine and Ravine of the Castles was southward and *up*valley are well founded. A significant post-icesheet episode of local glaciation in the ravines would have removed the imported mixed-lithology drift or deposited a drift mostly of domestic schist atop it. Bradley (1981, p. 324) suggests that a reestablished mountain glacier would take some time to remove a significant thickness of icesheet drift. But any such local ice is purely hypothetical, for it left no evidence such as local drift or looped alpine moraines.

Lower mountain flanks.—Much of the surficial material on the slope leading down to Randolph Valley from the upper north flank of the Presidential Range (fig. 10) is bouldery diamict, in places slightly hummocky, typical of ground moraine all over New England. Among the stones of Littleton Schist and micaceous gneiss are erratics of diverse lithology, including abundant granite, diorite, hornblende gneiss, and volcanic rocks, all of which crop out downvalley of the cirques (table 4; fig. 10). Most nonschist stones are subangular or subround, some are glacially faceted, some striated. Similar bouldery drift broadly mantles the lower northwest and west flanks of the range. Material on the north flank of the Presidential Range was deposited by a continental icesheet that transported material from the north or northwest. This drift lies not only downslope from King Ravine and Ravine of the Castles cirques but also downslope from the noncirqued Snyder Brook. On steeper slopes angular clasts of schist derived from bedrock inliers fan downslope across the mixed-lithology drift. On the lower slopes the drift sheet is irregularly cut by downslope-trending channels 1 to 3 m deep, some of which are outlined by marginal levees, which show that some of the drift has been reworked and redeposited as postglacial debris flows and stream alluvium. Mixed-lithology diamict irregularly laced by channels extends down the mountain flank to the floor of Randolph Valley.

Randolph Valley.—A diamict studded with south-derived schist boulders floors part of Randolph Valley. Bradley (1981) inferred this deposit to be till whose provenance indicates an advance of alpine glaciers after the icesheet disappeared—a sequence contrary to Goldthwait's (1970) inferred sequence. But Fowler (1984) argues convincingly that this diamict is colluvium, not till. Our 1983-84 field investigations, complete before the appearance of Fowler's paper of which we had no foreknowledge, generally corroborate the idea that much of the material is redeposited.

Landforms and exposures show that much of the Randolph Valley material is not till. A low-relief surface slope generally decreases from 5° to 3° from the lower mountain flank to the valley floor. Bradley's (1981) "only good exposure" of material he calls till at Durand Lake (fig. 10) shows 3 m of stratified silt, sand, and gravelly sand (fig. 11) in which

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Fig. 11. Stratified alluvium in Bradley's (1981) "only good exposure" of "till" at Durand Lake in Randolph valley, N.H. Alluvium is overlain by thin diamict containing boulders, probably a debris-flow deposit. Pen is 12.5 cm long.

planeset beds and foreset laminae dip northward downslope—clearly alluvial-fan material.

The bedded unit is capped by a diamict of gravelly mud about 0.5 m thick and studded with angular schist boulders as large as 3 m in intermediate diameter. Fowler (1984, table 1D) gives a stone-lithology count showing that at least 80 percent of the stones were derived from the Ammonoosuc Volcanic Series on the lower mountain flank and crystalline rocks from the north. Our lithology counts (table 4) suggest that 54 percent of stones in the diamict were derived from the Littleton Formation upslope to the south, and another 35 percent could be from the lower unit of the Littleton also upslope (table 4, fig. 10). Our lithology counts show greater percentages of stones of nonspecific origin (pegmatite, quartzite, vein quartz, undifferentiated mafics) than do either Bradley's (1981, p. 322) or Fowler's (1984, p. 435); we also recognize the lack of binary (Bickford) granite, which as Fowler notes should be present if local glaciers had moved northward from the cirques.

A reverse grading in the diamict is characteristic of the bases of particulate mass-flowage deposits such as pyroclastic flows (Sparks and others, 1973, fig. 2) and dense lahars (Pierson and Scott, 1985, fig. 10). The conventional interpretation of this phenomenon is that high shear stress and grain-dispersive forces at the base of a particulate mass flow tend to exclude coarse particles there (Bagnold, 1954). Recently a truly nutty explanation has emerged to answer an old riddle of why large Brazil nuts rise to the top of a jar of mixed nuts of like densities. During mixing (particulate mass flow) small particles can move by gravity down into the numerous developing transient voids that are too small to accept larger particles (Rosato and others, 1987). The reverse grading within the Durand Lake deposit suggests an origin as particulate mass flow (debris flow) rather than as till.

The diamict at Durand Lake thickens and coarsens up-fan (southward) to a railroad cut that exposes 5 m of poorly sorted, crudely stratified cobble gravel in a loose matrix of sandy granules. The clasts are mainly angular to subrounded (largely subangular) mica schist and gneiss; other clasts include granite, diorite, hornblende gneiss, quartzite, porphyritic basalt, pegmatite, and traces of red sandstone. Fowler (1984, table 1C) gives a stone-lithology count showing that at least 80 percent were derived from downslope or northern sources. Angular bolders of schist as large as 4 m in diameter were clearly derived from the upper Littleton Formation, which crops out on the upper mountain flanks ot the south (fig. 10) (Billings and others, 1979). But other boulders of granite and diorite, some of them striated, were originally derived from bedrock sources downslope or farther northwest. The lithology of the diamict thus reflects the upslope source: bedrock blanketed by mixed-lithology icesheet till. The noncompact, mud-poor nature of the material upslope along the railroad suggests that it is not a lodgment till.

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Schist stones are distributed not only downslope from the cirques (Ravine of the Castles, King Ravine, Bumpus Basin) but also downslope from the narrow, steep-walled, V-shaped, noncirqued Snyder Brook valley (fig. 10). Thus the downslope distribution of schist clasts is independent of the presence of cirques. Beneath mass-wastage debris from the steep sides of upper Snyder Brook valley, even the valley-floor drift contains metavolcanic and granite clasts imported from downvalley and from the north.

Large boulders like those in Randolph Valley also lie on valley floors near Crawford Notch and Pinkham Notch, deposited by historic alluvial torrents or debris flows (Morse, 1966, p. 82–95; Goldthwait, 1970b). Crandell and Waldron (1956), Beaty (1961), and Porter and Orombelli (1982) show how even experienced observers have mistaken debris-flow deposits for glacial drift. We concur with Gerath and Fowler and Fowler (1984) that the deposit at Durand Lake and the railroad cut upslope is a postglacial debris fan.

Absence of alpine moraines.—Neither aerial photographs nor field examination reveal moraines on the north flank of the Presidential Range. Bradley (1981) accounted the lack of moraines in Randolph Valley by speculating that local glaciers merged with a tongue of Laurentide ice flowing eastward along the valley. But this idea is inconsistent with the orientation of morainal ridges at the east end of the valley (fig. 10). The outer moraines are convex-south ridges indicating that the last ice in the eastern end of the valley flowed southward into Randolph Valley rather than northeastward from it. A railroad cut through the outer moraine exposes compact till having diverse granitic and gneissic stones, corroboration that the glacier was south flowing. Such a south flow into Randolph Valley is consistent with downwasting and northward recession of the icesheet in the Androscoggin Valley 3 to 25 km to the north (Gerath and others, 1985) and with Goldthwait's (1970b) inferred deglaciation of the region by downwasting and generally northward recession.

The Goldthwaits (1916; 1939, 1940, 1970a) have long reported a lack of alpine moraines in the Presidential Range and that such small subdued ridges that do exist are similar to countless chance ridges that the continental ice left all over New Hampshire. From air-photo and trailside examinations over the years in Tuckerman Ravine, Huntington Ravine, Gulf of Slides, Great Gulf, and other east-flank localities, we concur that numerous small ridges rich in angular local schist clasts are mass-wastage deposits such as protalus ramparts. We do not share Thompson's (1960, 1961) belief that this material buried earlier local-glacier moraines, for on some of the broader cirque floors the mass-wastage deposits are limited to relatively small areas at the bases of sheer walls.

On the lower west flank of the Presidential Range, sharp ridges lie on interfluves above the lower ends of the noncirqued Burt and Ammonoosuc Ravines (fig. 10). Burt and Ammonoosuc Ravines are cut into mica schist of the Littleton Formation, but the ridges consist of bouldery diamict that contains clasts of diverse granite, diorite, amphibolite, and gneissic stones, some as large as 1 m. The stones are rounded by distant transport; some are faceted or striated, and a few have pentagonal shapes considered as typical of till stones (Flint, 1971, p. 166–168). Included are clasts of Bickford Granite, biotite granite, and coarse granite, whose bedrock sources lie 3 to 12 km to the northwest (Billings and others, 1979). Nearby along the lower Cog Railroad where deforestation allows unobstructed views, similar drift continuously mantles a slope uninterrupted by moraines or gullies; this material is similar to the mixed-lithology icesheet drift that broadly mantles the west, north, and northeast flanks of the northern Presidential Range. Although the drift ridges descending steeply along and closely parallel to the incised Burt Creek contain mixed rock types, their positions are incongruous for icesheet moraines. The ridges probably are levees of huge debris flows that soon after deglaciation stripped loose icesheet drift off steeper slopes upvalley, descended Burt and other valleys, and spread out at the foot of the mountains. In these westside valleys we thus accept Thompson's (1960, 1961) proposal that the driftlike materials are mass-wastage deposits.

Conclusion.—As long advocated by J.W. Goldthwait and R.P. Goldthwait, the abundance of erratic stones on cirque floors and the absence of distinct alpine moraines indicate that the most recent substantial mountain glaciation in the Presidential Range cirques preceded the most recent icesheet occupation. Glacial drift and locally derived coarse clasts were redeposited downslope by postglacial debris flows and streams, not by post-icesheet mountain glaciers.

REGIONAL SEQUENCE OF DEGLACIATION AND MASS WASTING

Some workers in northern New England have suggested that the waning icesheet became segmented and locally stagnated as it thinned in high-relief terrain in west-central Maine (Borns and Calkin, 1977), near Mount Katahdin (Davis, 1976; Davis and Davis, 1980), near the Presidential Range (Gerath and others, 1985), and near the Green Mountains (Stewart, 1961, p. 23–26; 1971; Connally, 1968, 1971; Stewart and MacClintock, 1969, p. 44–45). Segmentation of downwasting continental ice by the emerging high-relief topography is a process by which the Cordilleran icesheet disappeared from the high-relief Washington Cascade Range (Waitt, 1972, p. 109–132; Waitt and Thorson, 1983, p. 64). Our studies in the northern Green Mountains, northern Presidential Range, and Mount Katahdin indicate or reaffirm that virtually all drift and depositional landforms there are deposits of the segmenting icesheet and subsequent mass wasting.

Goldthwait and Mickelson (1982) reviewed the evidence for downwasting and retreat of continental ice from New Hampshire in light of observed patterns of rapid deglaciation of high-relief terrain at Glacier Bay, Alaska during the last century. A small ice scrap orphaned from the thinning and retreating main glacier may locally reverse its flow direction for a few decades as the scrap downwastes. Recently deposited unvegetated drift lying on slopes is readily mobilized and redeposited downslope by mass-wastage and fluvial processes. Glacial debris is thus redeposited downslope even though ambient climate is melting existing ice masses and preventing formation of new glaciers. Continental ice downwasting and retreating from the highlands may strand bodies of ice thick enough in places to flow. Such ice scraps may briefly flow downslope enough to move small volumes of material but not enough to build recognizable moraines or erode cirques. The sequence of Pleistocene icesheet downwasting envisioned for the Presidential Range (Goldthwait, 1970a; Goldthwait and Mickelson, 1982) and the lack of a discrete episode of post-icesheet mountain glaciation are consistent with all field evidence known to us and reported in the literature from the Presidential Range and from other mountain areas of New England.

The melting of continental ice from New England highlands indicates that the regional snowline lay too high above the emergent highlands to sustain local alpine glaciers. Could a climate that melted existing ice build new glaciers at the same altitudes? Unlike higher parts of the Washington Cascade Range and numerous other high-alpine areas throughout the world, cirque floors in the New England mountains seem to have been too low even to intercept *slightly* lowered snowlines such as during the "Little Ice Age" of the past several centuries (Waitt, Yount, and Davis, 1982, fig. 6) or at any earlier time.

Many lower mountains flanks in the Green Mountains and Presidential Range display levees, channels, and other evidence that mountain drift has moved downslope in catastrophic debris flows or floods. Pollen evidence suggests that in northern New England slopes did not become stabilized by continuous forest until millennia after deglaciation (Davis and Jacobson, 1985, figs. 9–14). Rates of mass wasting of loose, unvegetated debris lying on slopes just after deglaciation may have been greater by an order of magnitude than they are now and probably remained high until forests became established. During the 13 millennia since deglaciation, moreover, rare earthquakes and high-energy storms doubtless have had cumulative influences on the valley floor that exceed those few events known from the two centuries of historic records.

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