Deglacial eolian regimes in New England

Robert M. Thorson*
Carol Ann Schile*
Department of Geology & Geophysics, 354 Mansfield Road, University of Connecticut, Storrs, Connecticut 06269

ABSTRACT

Ventifact-bearing cover sands, wind-driven lacustrine spits, and fluted bedrock outcrops in New England indicate that a strong anticyclonic circulation under moist periglacial conditions accompanied early recession of the Laurentide ice sheet. Dunes from this interval are conspicuously absent. An abrupt shift to warmer and drier conditions during the Bolling/Allerod period (12.7–11 ka) was apparently associated with rapid accretion of transverse dunes built by northeast winds along the shore of New England's largest ice-recessional lake. By the time of lake drainage, however, the anticyclonic circulation had been replaced by a northwesterly wind regime, resulting in erosion of the older deposits, the development of parabolic dunes, and gradual stabilization associated with revegetation. Reactivation of the dunes was associated with a return to colder and windier conditions during the Younger Dryas (11–10 ka) and with anthropogenic changes during the historic period, but not during Holocene time.

INTRODUCTION

Climatic changes associated with recession of the Laurentide ice sheet in northeastern North America have been simulated in separate climatic experiments (Manabe and Broccoli, 1985; Kutzbach and Guetter, 1986; Rind, 1987). In each case, the morphology of the Laurentide ice sheet, the areal extent of sea ice, and changing insolation patterns were largely responsible for determining model output (Fig. 1). In spite of large uncertainties associated with transient changes in the ice sheet (Clark, 1992), all of the models require the following conditions. (1) A subcontinental scale anticyclonic circulation must have been present over the ice sheet, and its intensity must have diminished proportionately with glacier recession. (2) Colder-than-present temperatures, especially during summer, and the alignment of storm tracks along the Atlantic margin must have contributed to moist periglacial conditions during early ice withdrawal. (3) The breakdown of the glacial anticyclone would have allowed westerly airflow to resume.

Can these predictions be confirmed? The abundance of sedge and willow in the early pollen records (Davis, 1983; Webb et al., 1987) is consistent with simulations requiring colder and wetter conditions during the early part of the deglacial hemicycle (18–14 ka [1 ka = 1000 ^{14}C yr B.P]; COHMAP, 1988). Although the band of forb pollen south of the ice sheet at 12 ka suggests that the ice sheet had a peripheral effect on plant communities, the regional pollen stratigraphy cannot be used to confirm the direction, intensity, or seasonality of the wind regime. Other proxy climatic records (tree-line variations, lake records, and insect macrofossils) are also not directly dependent on the paleowind regime and are generally restricted to the Holocene epoch. The localized, transient development of permafrost during glacier recession is suggested (Stone and Ashley, 1992) but not confirmed (Black, 1983) by the geologic evidence, and it provides no useful constraint on the paleowind regime.

In contrast to other proxy records, the eolian deposits of New England can be used to confirm some results of these climate simulations and to evaluate the terrestrial response to abrupt climate changes, because, unlike biological data, eolian deposits define limits for the direction, seasonality, and strength of former winds (Sweet, 1992), parameters integral to the design of atmospheric general circulation models (GCMs). Eolian bed forms (dunes and ripples) are especially valuable because their primary sedimentary structures reflect specific, instantaneous threshold conditions, rather than the integrated effects of biological and hydrophysical processes. For example, David (1981) used the eolian stratigraphy in the northern Great Plains to demonstrate not only the existence of the glacial anticyclone along the western sector of the Laurentide ice sheet during early Holocene time, but an abrupt reversal in the prevailing direction of sand-moving winds from southeasterly to westerly associated with its breakdown. Recent reconstructions in Alaska (Lea and Waythomas, 1990), the Pacific northwest (Barnosky et al., 1987), and in the north-central United States (Forman

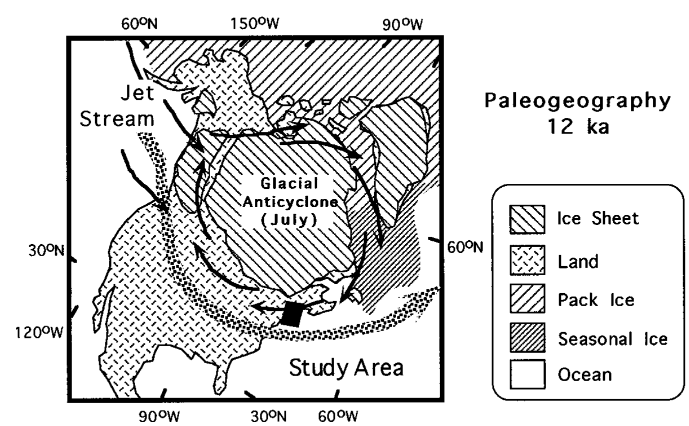

Figure 1. Environmental conditions at 12 ka showing the location of the study area relative to the anticyclonic circulation required by the presence of the Laurentide ice sheet. Adapted from COHMAP (1988).

*E-mail address (Thorson): thorson@geol.uconn.edu. Present address (Schile): Miller Environmental Management, Inc., P.O. Box 455, Grayslake, Illinois 60030-0455.

GSA Bulletin; July 1995; v. 107; no. 7; p. 751–761; 11 figures; 1 table.

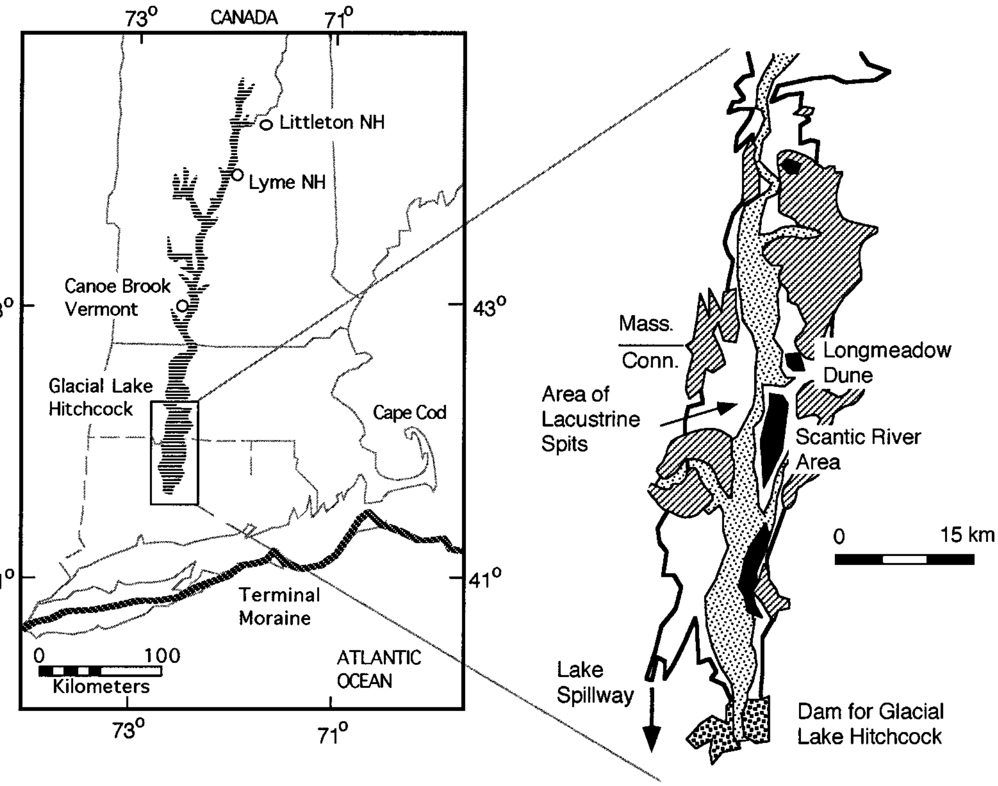

Figure 2. Map showing extent of Glacial Lake Hitchcock (from Ridge and Larsen, 1990) and location of terminal moraine (from Stone and Borns, 1986). Inset map (adapted from Stone and Ashley, 1992) shows geologic relation between dunes (black) and sediments associated with Glacial Lake Hitchcock (lake-bottom sediments, white; meltwater deltas, striped; post-lake terraces, light stipple; drift dam, coarse stipple).

et al., 1992) demonstrate the unique value of eolian bed forms in confirming numerically simulated paleowind regimes.

Although late Pleistocene eolian sediments are ubiquitous throughout New England, they have received little attention at the regional scale, and their value in testing paleoclimatic simulations continues to be undervalued (Kutzbach and Wright, 1985). This paper reviews the sparse, qualitative literature on the eolian deposits for New England, interprets new evidence for the paleowind regime associated with spit formation in Glacial Lake Hitchcock, and describes the sedimentology of what may be New England's largest dune. We integrate our findings into a conceptual model for changing eolian regimes during the deglaciation of New England.

REGIONAL SETTING

The overall style of deglaciation for the New England sector of the Laurentide ice sheet has been known since the late 19th century (Fuller, 1914; Schafer and Hartshorn, 1965). Recent reviews by Mickelson et al. (1983), Gustavson and Boothroyd (1987), Stone and Borns (1986), Ridge and Larsen (1990), Lewis and Radway-Stone (1991), and Stone and Ashley (1992) demonstrate that recession from the terminal moraines in the coastal lowlands (Fig. 3) probably began about 20–22 ka. After constructing a series of recessional moraines near the ice limit prior to about 17 ka, the glacier withdrew rapidly northward, reaching the St. Lawrence River about 12 ka (Karrow and Occhietti, 1989). Collectively, these records indicate that a prolonged period of thinning and slow recession under periglacial conditions was followed by rapid ablation and accelerating retreat (Fig. 3).

The deglacial chronology for western and central New England is tied to Glacial Lake Hitchcock (Lougee, 1939), the largest and most long-lived of New England's ice-recessional lakes (Schafer and Hartshorn, 1965; Koteff et al., 1988; Fig. 2). Glacial Lake Hitchcock

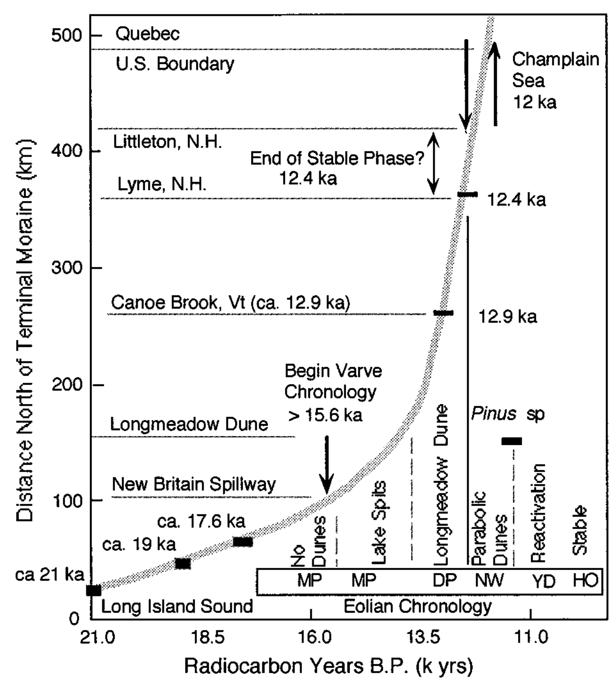

Figure 3. Time-distance diagram showing recession of the New England sector of the Laurentide ice sheet drawn parallel to the eastern boundary of Glacial Lake Hitchcock in Connecticut and Massachusetts from Figure 2. Older dates are from Stone and Borns (1986); younger dates are from Ridge and Larsen (1990) and Karrow and Occhietti (1989). Also shown is the sequence of eolian features and regimes described in the text (MP, moist periglacial; DP, dry periglacial; NW, northwesterly; YD, Younger Dryas; HO, Holocene).

formed when drainage from the Connecticut River watershed was blocked by glaciolacustrine deltas near Middletown, Connecticut, and routed through a bedrock-floored spillway to the west. Glacial Lake Hitchcock expanded northward as the ice withdrew, reaching northern New Hampshire before the drift dam failed. As the lake grew northward, emergent, high-level ice-contact deltas and lake-bottom deposits were incised and exposed to eolian processes. Although this regression is generally attributed to spillway erosion, isostatic rebound with a fixed outlet provides an equally viable explanation.

The 4000-yr chronology of Glacial Lake Hitchcock originally established by Antevs (1922) is now linked to the radiocarbon record (Ridge and Larsen, 1990). Dates obtained on detrital wood fragments from lake-bottom varves ". . . place the inception of Lake Hitchcock in central Connecticut at before 15 600 yr B.P., and deglaciation of [southern Vermont] at about 12 900 yr B.P. An abrupt change in sediment types and thickness of varves . . . about 12 400 yr B.P. . . . corresponds to a basin-wide change in the New England varve chronology and records the initial breaching of the dam for Lake Hitchcock" (Ridge and Larsen, 1991, p. 889). We adopt their date of 12.4 ka (14 510; 14 718–14 315 calendar yr B.P.) as a general date for lake drainage.

Stone and Ashley (1992) recently suggested that the Connecticut portion of Glacial Lake Hitchcock drained as early as about 14 ka, an interpretation based on accelerator mass spectrometry (AMS) dates ranging from 13.5 to 13.7 ka on reworked plant detritus collected from a delta lobe deposited during shoreline regression. Their interpretation conflicts with the direct calibration of the varve record by Ridge and Larsen (1990) and with a suite of basal peat dates obtained from small depressions of the floor of Glacial Lake Hitchcock interpreted as the scars of pingos that formed during exposure of the lake floor. Although a "pingo-scar" origin for the depressions is debatable, their chronology is not; basal radiocarbon dates from four of the depressions range from 11 890 $\pm$ 130 to 12 630 $\pm$ 240 ^{14}C yr B.P.,[1] indicating near simultaneous formation shortly after a period of eolian deposition on the exposed lake floor. Controversy continues because the basal peat dates could indicate simultaneous formation of the depressions, rather than lake-floor exposure, and because the dates on lake drainage to the north may not be relevant to the timing of drainage to the south. The simplest explanation, however, is that one lake drained shortly before the basal peat deposition.

A REVIEW OF EOLIAN FEATURES

The Dearth of Dunes

Although the eolian deposits of New England are easily recognized and nearly ubiquitous, especially in lowlands dominated by meltwater facies, the dearth of diagnostic sedimentary structures precludes meaningful comparisons with other periglacial eolian deposits (e.g., Schwan, 1988; Lea, 1990). Since the 1930s, when the eolian, rather than pedogenic, origin of the sediments was demonstrated (e.g., Denny, 1936), these deposits have been informally referred to as the "eolian mantle" (Flint, 1930) and largely ignored in regional investigations. Over most of its distribution, the eolian mantle is too coarse and poorly sorted to be called loess, but too thin (<1 m) and locally discontinuous to be called "cover sand." Sediment texture ranges from moderately sorted fine sandy loams to loamy sands (0.07–0.25 mm). The deposits typically lack sedimentary structures (Fuller, 1914) and mineralogical differences (Smith and Fraser, 1935) that would permit the wind direction or source areas to be defined, respectively. Additionally, they are extensively bioturbated and texturally fractionated by pedogenic processes (Johnson, 1990), preventing differentiation into meaningful stratigraphic units, and limiting opportunities for radiocarbon and thermoluminescence dating.

Left unsaid in the cursory treatment of the eolian mantle is the *conspicuous absence* of sand dunes and bedded sand sheets throughout most of the region, the *conspicuous absence* of extralocal trends in thickness or texture, and the *conspicuous absence* of loess. In other words, the negative evidence against regional deglacial aridity has not been emphasized.

The absence of dunes is particularly notable in New England's southeast coastal zone, where edaphic conditions favorable for dune formation (broad lowlands, braided sandurs, katabatic outflow) were present prior to revegetation (Mather et al., 1942, p. 1163; Bryan, 1931). Based on the abundance of ventifacts, and the absence of an eolian fill in kettles, Mather et al. (1942) concluded that sand movement on Cape Cod was extensive but that it occurred early, and under a cold periglacial climate. Fuller (1914, p. 212) reported that ". . . there is little evidence of dune formation [on Long Island], notwithstanding the favorable character of the materials. . . ." Hartshorn (1967) used textural analysis to demonstrate that traction (saltation) was the dominant transport mechanism for sand sheets in southeastern Massachusetts, even though surface morphology and sedimentary structures were absent. The leeward equivalent of these deposits lies to the west in the Boston Basin, where the eolian mantle has a higher silt content (Smith and Fraser, 1935).

Eolian deposits elsewhere in New England are best developed on the deltaic sand plains associated with local concentrations of meltwater deposits (e.g., Clebnik, 1980) and with regression from the marine limit north of Boston (McKeon, 1989). In such settings, the sand supply was abundant, surfaces of low relief were exposed to dune-forming winds, and vegetation must have been minimal. Nevertheless, eolian deposits in these areas also lack diagnostic structures, and dune morphology is restricted to incipient longitudinal forms.

As one of the few radiocarbon-dated exposures of the eolian mantle, Thorson and McBride (1988) described the eolian stratigraphy in the leeward wind shadow of a prominent bedrock ridge just east of Glacial Lake Hitchcock. At that site, postglacial soils began to develop on deposits characteristic of the eolian mantle sometime before 10.6 ka, perhaps during moist conditions associated with the Younger Dryas climatic reversal (10–11 ka). However, a distinctly younger, much finer grained deposit resembling loess was deposited during early Holocene time (10.6–8 ka), when southern New England was clearly drier than present (Davis, 1983; Webb et al., 1993). This relationship suggests that the absence of fine-grained eolian sediments from earlier time periods resulted from conditions too moist for the effective entrainment and transport of dust. No source for the early Holocene "loess" has yet been found.

Ventifacts and Fluted Bedrock

Ventifacts are widely reported throughout New England (Bryan, 1931; Schafer and Hartshorn, 1965), but seldom are they well documented. Only on the outwash plains near the terminal zone are they well developed (faceted) or abundant (Mather et al.,

[1]Other dates include 12 050 $\pm$ 110, 12 200 $\pm$ 250, and 12 300 $\pm$ 110 ^{14}C yr B.P., and a detrital wood fragment dating to 14 330 $\pm$ 430 and thought to have been reworked from older deposits.

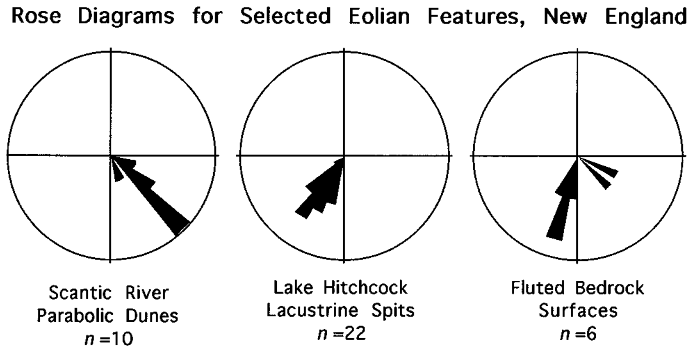

Figure 4. Rose diagrams showing directional data for selected eolian features. Methods for obtaining data for Scantic River dunes and Lake Hitchcock spits are described in text. Data for fluted bedrock surfaces are from Hartshorn (1962). Circle diameter is scaled to represent 40% of total number of observations (*n*).

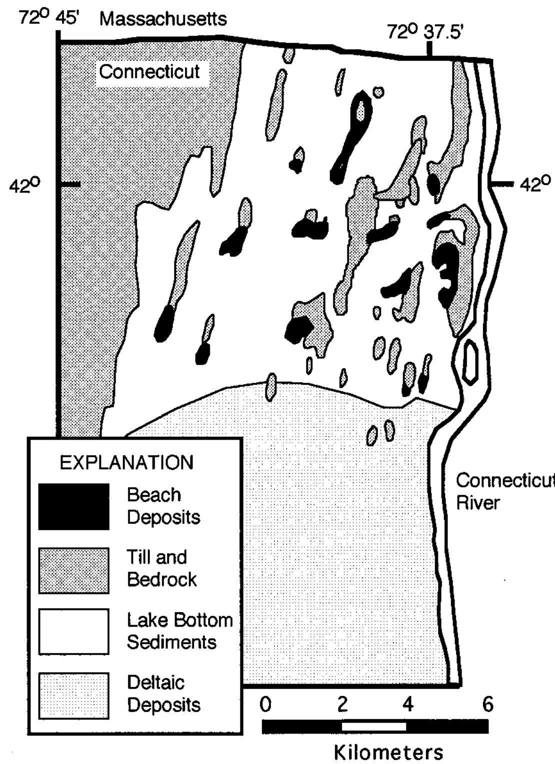

Figure 5. Lacustrine spits redrawn from U.S. Geological Survey quadrangle maps (Cushman, 1963; Colton, 1960; Hartshorn and Colton, 1971) onto simplified geologic map adapted from the state materials map (Stone and Schafer, 1992).

1942). The presence of ventifacts within outwash sequences near the ice margin suggests that they formed in settings proximal to the ice margin (Oldale and O'Hara, 1984). Ventifacts elsewhere in New England typically consist of slightly polished, unfaceted stones with a sporadic local distribution (McKeon, 1989). Experimental studies indicate that the rate of sandblasting leading to ventifact formation is limited not by the wind strength or soil moisture, but by the availability of saltating sand (Sharp, 1980). Thus, the limited ventifact formation north and west of the coastal zone suggests the restricted mobility of sand during ice withdrawal.

Although fluted bedrock surfaces and boulders have been reported for more than a century, Hartshorn (1962) was the only investigator to inventory them or to evaluate them for their paleoclimatic significance. Fluted bedrock surfaces, like ventifacts, are largely restricted to southeastern New England. Of the ten fluted surfaces reported (Schafer and Hartshorn, 1965), ". . . eight were cut by winds from the north or north-northeast; only two were cut by winds from the northwest or west." The small population of flutes reported in Hartshorn's earlier analysis (1962) yields a bimodal distribution (Fig. 4 and Table 1). The predominance of flutes indicating northeasterly winds suggests that they were associated with ice-marginal anticyclonic winds.

Lacustrine Spits

Cushman (1963) and Colton (1962) reported the presence of gravelly beach deposits that formed at and below the strandlines of Glacial Lake Hitchcock in northern Connecticut and southern Massachusetts (Fig. 5). These deposits project beyond the southern ends of drumlinoid hills that stood as islands within the lake and take the form of "southwestward-curving spits and bars." Although their existence was noted and mapped by Schafer and Hartshorn (1965) and Stone and Schafer (1992), the eolian significance of these features has only recently been noted by Janet Stone (1994, written commun.).

We examined the stratigraphic setting, morphology, and eolian implications of these deposits as mapped by Cushman (1963) and Colton (1960). Our qualitative analysis provides the following information. (1) The spits extend in a direction only to the southwest of islands in the lake. (2) They have a simple morphology with an apparent concave-northward curvature, but with no systematic trends in thickness, slope, or height. (3) There is no apparent re-

TABLE 1. DIRECTIONAL DATA FOR EOLIAN FEATURES IN NEW ENGLAND

Feature	Dominant structures	Sample size (n)	Azimuth of trend		Dip magnitude	
			Mean	S.D.*	Mean	S.D.
Cross-bedding, Longmeadow Dune						
Grain flow	Unbedded; flow bands	18	206.5	14.7	28.8	4.2
Grain fall	Nondistinct graded and graded	159	195.8	23.5	13.7	7.8
Ripples	Inverse graded and ripple forms	93	181.7	131.7	5.0	3.0
Plane bed	Nondistinct graded	8	36.9	12.8	4.2	1.4
Truncation	Stratigraphic discontinuity	41	211.2	80.3	7.8	5.7
Lacustrine spits, Glacial Lake Hitchcock		22	208.2	13.1		
Parabolic dunes, Scantic River area		10	311.8	17.9		
Fluted bedrock surfaces, eastern Massachusetts		6	163.0	153.2		

*S.D. = standard deviation.

lationship between the size of the island and the apparent volume of the beach deposits. (4) The spits appear to be restricted to the widest part of Glacial Lake Hitchcock, but there does not appear to be a relationship between local wave fetch and degree of development. (5) There is a well-defined upper limit to the mapped distribution of beach deposits (Cushman, 1963). (6) The beach deposits extend continuously downward from an elevation of about 60 m (200 ft) to the floor of the paleolake at about 30 m. (7) There are no secondary spits developed on the larger and older ones.

These relationships indicate that the beach deposits were built by littoral currents associated with waves on the surface of an ice-free (summer/autumn) lake. The consistent location of the spits indicates that they formed by the net accumulation of storm-eroded material in protected leeward settings. The apparent curvature to the spits suggests an easterly component for wind-driven waves. The continuous range in elevation of the spits and their consistent orientation on different shoreline positions of Glacial Lake Hitchcock indicate that a southwesterly leeward wind shadow remained fixed during spit formation, most of which occurred early in the lake history.

We collected directional data on spit elongation for the most highly concentrated and well-mapped group of spits in the Lake Hitchcock basin. We measured the azimuth of 22 spits by bisecting a triangle drawn from the apex of each drumlinoid hill to the western and eastern borders of each spit (Fig. 4). Spit azimuth ranged from 185° to 233°, with a mean of 208° ± 13°. On the basis of this data, we conclude that strong summer winds driven by the glacial anticyclone were responsible for creating the spits, a phenomenon that was especially important during the early phase of the lake.

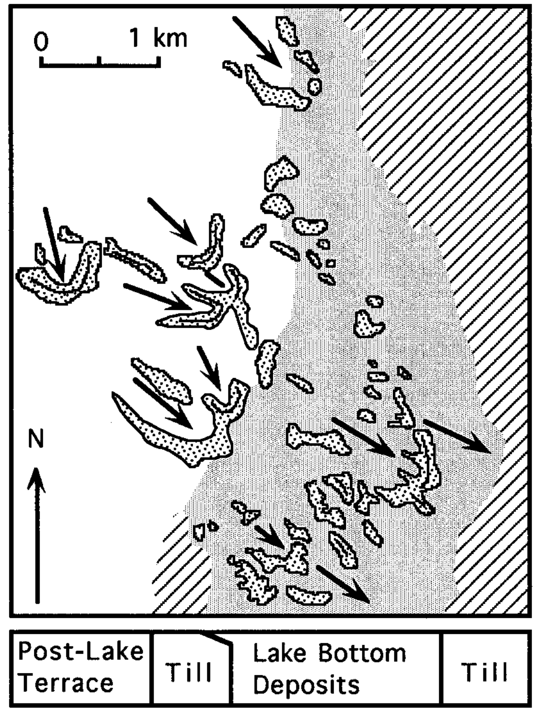

Figure 6. Incipient parabolic dunes in the Scantic River area, developed on post-lake terraces and lake bottom deposits. Arrows indicate inferred wind direction responsible for dune formation. After Colton (1965b) and Stone and Ashley (1992).

Parabolic Dunes

Incipient dunes are widespread in the deglaciated parts of northeastern North America. Parabolic dunes in the Hudson River Valley were built by northwesterly winds following the drainage of Lake Albany (Dineen et al., 1988). Dunes associated with glacial lakes in the eastern Great Lakes (Karrow and Occhietti, 1989) and with the marine incursion in the St. Lawrence Valley (David, 1989; Filion, 1987) are also generally parabolic in form and were built primarily by northwesterly winds. In west-central Maine, longitudinal dunes developed under a unimodal northwesterly wind regime during marine regression (McKeon, 1989).

Conspicuous parabolic dunes of late Pleistocene age are associated with Lake Hitchcock (Jahns and Willard, 1942). In spite of their size (up to 2 km long and 15 m high) and unambiguous morphology (Cushman, 1963; Colton, 1965a, 1965b; Hartshorn and Koteff, 1967; Hartshorn and Colton, 1971), little has been written about the dunes, and no quantitative sedimentological analysis was published prior to Schile's (1991) investigation of the stationary transverse dune at Longmeadow. Hartshorn (1962) was the first to recognize that the dune field at Longmeadow was built by north-northeasterly winds and formed under a wind regime distinctly older than the parabolic dunes that developed on younger surfaces (Fig. 6).

Parabolic dunes formed on the lake-bottom sediments and post-lake terraces of Glacial Lake Hitchcock have the following characteristics. (1) They are most common on the eastern side of the Connecticut Valley, a persistent pattern as far north as Vermont. (2) Although a sand sheet of variable thickness occurs widely across the lake-bottom and deltaic deposits, distinct dune morphology is only locally present. (3) Primary sedimentary structures are rare, being exposed only where deposits are thicker than several meters (we are aware of only one published reference to measured bed inclination; 7°–33° to the southeast; Hartshorn, 1962). (4) Southeasterly oriented parabolic dunes up to 1 km long and 9 m high are the only dune type that can be confidently identified; elongate sand ridges of unknown origin also occur widely within and beyond the clusters of parabolic dunes. (5) No dunes convincingly associated with northeasterly winds are known from either the floor of Glacial Lake Hitchcock or younger terrace deposits.

These observations have been correctly interpreted by previous investigators to indicate that northwesterly winds were responsible for deflation of the lake-bottom sediments, the net southeasterly transport of sand, and the formation of parabolic dunes. The general absence of other dune types, notably barchanoid, linear, and transverse dunes, and the dearth of primary structures suggest that the final phase of dune growth occurred in the presence of vegetation. The overall dune shape, low length-width ratios and broad but irregular leeward slopes, suggests only incipient development prior to stabilization (McKee, 1979). A limited sample of these parabolic dunes (n = 10) from the Scantic River area, where they are best defined, yields a mean paleowind direction of 311.8° ± 17.9° (Fig. 4; Table 1).

THE LONGMEADOW DUNE

Morphology and Methods

The Longmeadow dune is the northernmost of two elongate ridges that occupy the southern, downwind edge of a dune field near the eastern shoreline of Glacial Lake Hitchcock, about 2 km from the till-covered upland to the east (Fig. 2). The remainder of the field consists of irregular sand mounds and a single well-defined northeast-trending parabolic dune. This dune field occurs on the tread of a broad (2 km wide) terrace that is underlain by lake-bottom deposits associated with the highest shoreline of Lake Hitchcock, and 3–6 m above the "stable" shoreline level to the west (Stone and Ashley, 1992).

The external topographic form of the Longmeadow dune is an irregular ridge about 10 m high and 1 km long, that trends to the east-southeast with an azimuth of about 100° (Fig. 7). In its central section, where the dune is highest, the ridge is asymmetric, with the steeper side facing due south, and with a ridge-crest slightly concave to the south (Fig. 8). These morphological attributes are consistent with the internal stratigraphy of the dune, which is dominated by windward and leeward facies on the north and south sides of the asymmetric ridge, respectively. Although the dune's external form appears to have been modified by younger secondary winds from the northwest, the primary morphology of the dune is that of a transverse dune or barchanoid ridge (McKee, 1979), a form characteristic of strongly unimodal winds with an abundant sand supply (Freyberger, 1979).

Between 1988 and 1992, the internal stratigraphy of the dune crest was intermittently exposed during the extraction of sand prior to development of the site for suburban housing. During that time, we measured five stratigraphic sections showing the internal features of the dune, ranging from 5 m to 102 m in length (Fig. 7). Our terminology for cross-bedding and eolian stratification follow McKee and Weir (1953) and Hunter (1977), respectively. Terminology for wind direction relative to dune form follows Sweet (1992). Depositional facies follow those of Lea (1990). Paleocurrent data were measured by the bed-intersection method. Samples for grain-size analyses were obtained from exposed faces, sieved at quarter phi intervals, and plotted using the procedures of Folk (1968).

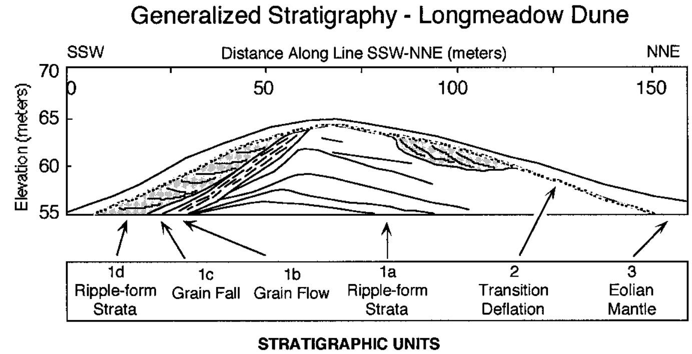

Figure 8. Generalized stratigraphic section of the Longmeadow dune perpendicular to its crest. Stratigraphic units are described in text.

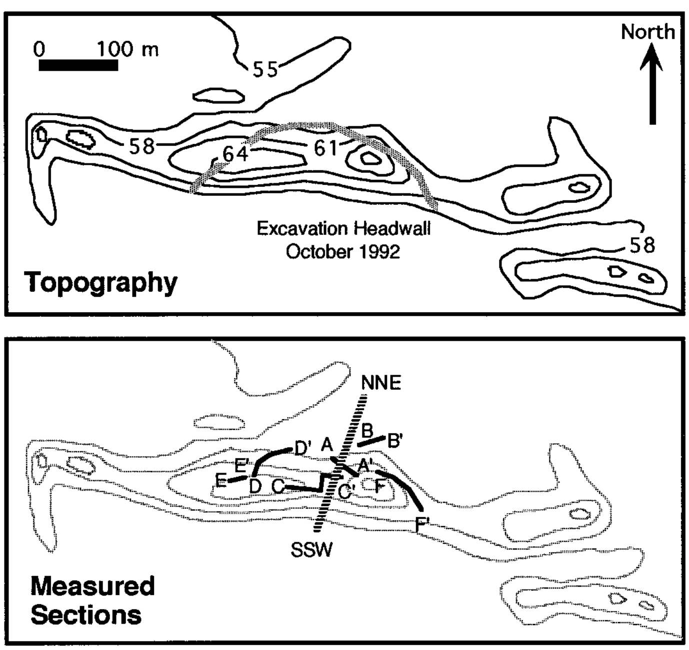

Figure 7. Maps of the Longmeadow dune showing the topography based on the U.S. Geological Survey topographic map of the Springfield South quadrangle, the maximum extent of the headwall excavation, and the location of measured sections described in the text. Line labeled SSW-NNE shows location of section shown in Figure 8.

Stratigraphy

The Longmeadow dune consists of three principal stratigraphic units, from oldest to youngest (Figs. 8 and 9). Unit 1, which is up to 11 m in exposed thickness and forms the bulk of the dune, is dominated by unbioturbated, laminated and tabular-planar cross-bedded eolian sand that accumulated during the upward growth of the dune. They are consistently better sorted and more clearly bedded than the overlying units. Unit 2 is a transitional deposit ranging from 0.25 to 1.4 m in thickness that sharply overlies the well-bedded dune sands along a continuous erosional unconformity. It is dominated by lag concentrations of granules and coarse sand on broadly curved truncation surfaces (blowouts), and by intergraded medium-fine sand and fine sand. Unit 3 conformably overlies the transitional zone and is distinguished from it by the absence of both granules and primary sedimentary structures, by its elevated silt content, and by its browner, more thoroughly oxidized color. It mantles the dune as a continuous drape that thickens toward the base of the dune.

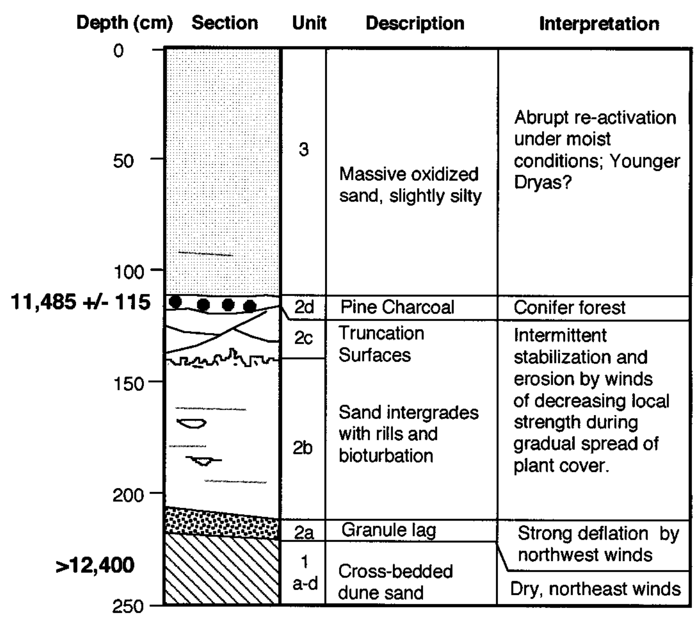

Figure 9. Section E–E showing subdivision of eolian strata near the dune crest. See text for description.

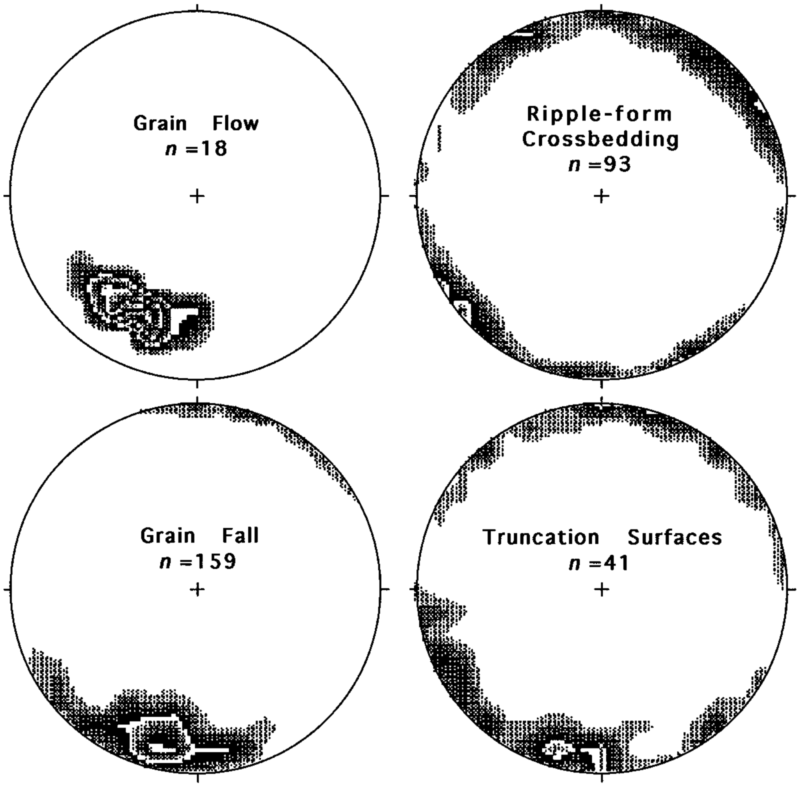

Figure 10. Lower-hemisphere stereo plot (1% equal area contours) showing dip azimuth and dip angle of selected eolian bed forms in the Longmeadow dune. Sedimentary structures are based on Hunter (1977).

Primary Sedimentary Structures. The Longmeadow dune exhibits all of the primary sedimentary structures characteristically present in modern inland dunes characteristic of arid areas (Table 1 of Hunter, 1977). These include erosional truncation surfaces (coarse-grained lag horizons), tractional deposits formed under conditions of attached flow (ripple-form strata and plane-bed lamination), sediments deposited from suspension on the leeward face of the dune (grain-fall strata), and sediments associated with dry avalanching down the leeward dune face (grain-flow strata). The most common bed form is inversely graded, subcritically climbing, translatent strata, which develop during conditions of net deposition associated with migration of wind ripples on a dry unvegetated surface. Stratigraphic relations for the five sections described individually by Schile (1991) are generalized into a single composite section perpendicular to its crest in the following discussion.

The Dune Sand: Unit 1. Five meters of large-scale tabular-planar and wedge-shaped cross-bed sets were exposed on the northern (stoss) side of the dune (A–A′). Laterally continuous inversely graded translatent laminae dipping gently to the northeast (Fig. 10; mean azimuth = 24°) dominate the section. Nondistinct graded stratification interpreted as plane-bed deposition, ripple-form laminae with wave-form spacing of 40–50 cm, and lag horizons extending continuously across the exposure face are locally abundant. These structures indicate deposition on the windward slope of an accreting dune by attached, unimodal flow on a dry, unvegetated surface. Using the threshold drag velocity as calculated by Freyberger (1979), summer winds at Longmeadow must have frequently exceeded 6 m/s. Short-term rates of net deposition ranged from strongly positive (burial of coarse-grained intact ripples) to negative (deflation) values. Exposures nearest the underlying lacustrine strata (section B–B′) are also dominated by inversely graded translatent strata that are inclined to the southwest (234°), rather than the northeast, suggesting the incipient upward growth of a sand mound (ribar dune) during attached flow. The abundance of coarse-grained lag horizons, plane-bed lamination, and vertical stacking of preserved ripples suggest an oscillating sediment supply and very strong winds.

The longest exposure (section C–C′) extends a distance of 102 m subparallel to the dune crest. In the eastern subsection, the deposits are dominated by nearly horizontal cross-strata, composed of tabular-planar and simple low-angle cross-bed sets, 12–15 cm thick, dominated by ripple-form strata. All of the strata are inclined <7° and exhibit a range of bedding orientations to the northwest, northeast, and southwest. The deposits represent the migration of wind ripples on an accreting dune, with the sand transport becoming more complex with increasing dune height. The central part of section C–C′, cut at nearly right angles to the dune crest, records the transition from conditions of attached flow on the windward side to conditions of flow separation over the brink of the dune and permits the subdivision of unit 1 into meaningful subunits (units 1a–1d). The ripple-form strata near the top of the dune on its windward side (unit 1a) can be traced continuously southward into steeply inclined (22°–34°) grain-flow strata (unit 1b) that trend consistently to the southwest (206° ± 15°). Over most of the section, the grain-flow strata are overlain by grain-fall laminae (unit 1c), but near the base of section, ripple-form strata (unit 1a) are unconformably overlain by grain-fall and grain-flow deposits. The western part of section C–C′ exposes a thick sequence of grain-fall strata above the grain-flow deposits that dip less steeply upward. This suggests that the interval associated with peak flow separation, which led to the steep slope responsible for grain-flow strata, was brief. The grain-fall strata are interbedded with ripple-form strata, indicating that the angle of incidence of winds relative to the crest shifted irregularly in time. Unit 1d, exposed only along the western edge of section C–C′, lies conformably above the grain-fall deposits (unit 1c). It is distinctly coarser in texture, is less clearly bedded, and is dominated by inversely graded climbing ripple cross-laminae, which record a variety of bed inclinations, with a mean northwesterly azimuth of 287°.

The western part of the dune (D–D′) is dominated by normally graded beds interpreted as leeward grain-fall deposits that overlie southwest-dipping (18°–28°) grain-flow laminae. Near the top of the section, however, the grain-fall laminae dip 8°–16° to the northeast above a smoothly curved unconformity. This reversal of bedding suggests that the final deposition occurred as sand was blown over the dune crest with a westerly or southerly component of flow.

The Transition Zone: Unit 2. A distinct transition zone, ranging from 25 to 140 cm thick, lies unconformably above the dune sands on both stoss and lee slopes, yet conformably beneath the overlying eolian mantle (E–E′; Fig. 9). The base of the transition (unit 2a) is a lag horizon consisting of one or more diffuse concentrations of coarse sand and granules between 2 and 15 cm thick. Next in sequence (unit 2b) is a horizon texturally similar to the underlying dune, but stratification is manifest only as a faint banding; shallow cross-bedded rill deposits and small burrow casts are also present. Unit 2c consists of a strongly turbated mottled oxidized coarse sand of an incipient paleosol unconformably overlain by lenticular cross-beds of massive sand with broadly curved basal truncation surfaces. Unit 2d is a charcoal-rich horizon present in several exposures just below the top of the transitional horizon. Charcoal fragments from *Pinus* sp. are dispersed throughout a massive sand matrix. A small fragment of this charcoal yielded an accelerator ^{14}C date of 11 485 ± 115 yr B.P. (13 400; 13 554–13 265 calibrated yr B.P. AA-7154).[2]

Eolian Mantle: Unit 3. The eolian mantle ranges from 60 to 140 cm in thickness. Although similar in bulk texture to the underlying dune sand, unit 3 is distinguished by its stronger soil color (10YR 5/6-8) irrespective of depth below the modern solum, by its greater silt content, and by the complete absence of bedding and primary and sedimentary structures. The modern solum consists of a 5–8 cm thick ochric epipedon (A horizon), overlying a structureless but highly oxidized incipient B_w horizon. Much of the eolian mantle lies below the root zone, suggesting that its uniform color does not result from Holocene pedogenesis. Locally, the Holocene solum is buried by as much as 30 cm of massive loamy sand, which in turn lies beneath the modern forest floor and humus layer.

Secondary Sedimentary Structures. Secondary structures occur widely in the upper 1–2 m of the primary dune strata (unit 1). We interpret locally massive graded laminae with sharp basal contacts near the base of the dune as adhesion structures (Kocurek and Fielder, 1982) resulting from the deposition of sand across a wet surface. Other evidence for surface moisture includes concentrations of oxidized silt which crosscut and postdate near-surface depositional layering (silty infiltration structures; McKee and Bigarella, 1972). Enhanced cohesion due to moisture on the leeward side of the dune is indicated by break-apart laminae (local extension), by en echelon faults associated with slumping, and by isolated and rotated blocks of sand in the grain-flow deposits. Transported blocks within grain-flow laminae have rounded edges and are restricted to proximal leeward slopes, and the en echelon faults have offsets and spacing consistently <5 cm. These observations suggest that the dune strata had negligible bulk strength prior to failure (unfrozen). Localized and linear concave-upward disturbances, which crosscut the underlying laminae, are interpreted as ephemeral rills, the sides of which collapsed in a saturated state during their filling.

In the more gently dipping tabular cross-bed sets of unit 1c, silty infiltration structures in the grain-fall laminae are locally fragmented and offset slightly. Similar features were attributed to localized freeze-thaw processes by Ahlbrandt and Andrews (1978). Above this unit is an enigmatic tabular cross-bed set in which the grain-fall laminae are deformed into similar folds with an amplitude of 7–10 cm and a spacing of nearly 1 m. Axial planes in the folds are parallel to the leeward slope of the dune and overturned to the southwest. We interpret this phenomenon as ductile folding that occurred under conditions of longitudinal compression set up by a shallow translational slide down the leeward slope of the dune. Movement in a saturated, but unfrozen state is suggested, a condition that suggests rapid thawing of seasonal frost on the southwest-facing slope.

Bioturbation structures were restricted to the upper 2 m of the dune deposits (unit 1) and were much more common on the south-facing leeward slope. Vertical-walled chambers between 2 and 10 cm high and between 1 and 4 cm wide sharply crosscut the ripple-form strata on the windward face and are interpreted as animal (insect?) burrows. Similar structures on the leeward face appeared as horizontal chambers above coarse-grained bedding contacts that were connected by vertical tunnels between laminae. The size, location, and morphology of these structures indicate that they result from the burrowing activities of insects (Howard, 1975), primarily on the warmest and most well-drained (southwest) slope of the dune. We found no burrows larger than 4 cm in diameter, and no animal trackways.

Branching cylindrical-shaped intrusive structures are present in the upper meter of the dune and increase in frequency upward. They are interpreted as root casts associated with former vegetation. One notable exception to this rule is a 10-cm-thick well-defined zone of intense oxide mottling that crosscuts grain-flow laminae near the top of the leeward face and is buried by additional grain-flow deposits. The mottles increase in frequency upward to the middle of the horizon, where they merge to form a continuous band that is sharply overlain by dune deposits. This feature represents an incipient paleosol that formed below shallow-rooted herbaceous plants and represents a temporary pause in dune accretion.

Granulometry. Based on an analysis of fifteen samples sieved at quarter phi intervals (Fig. 11), the material texture in the Longmeadow dune spans the range from granule gravel to very coarse

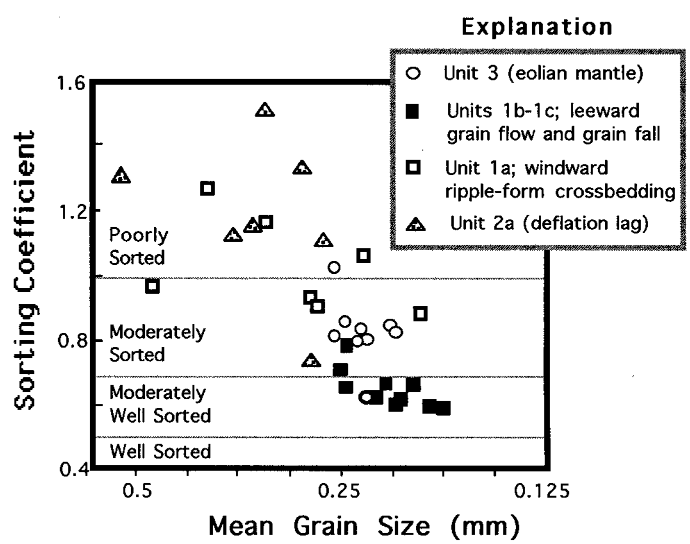

Figure 11. Textural data from the Longmeadow dune. See text for explanation.

[2]Identified and dated by Lucinda McWeeney.

silt, with a mean value of 0.25 mm (1.97 phi). Ahlbrandt (1979) reported a nearly identical mean value of 1.96 phi for a representative sample of inland dunes in western north America. Deposits from the Longmeadow dune, however, are more poorly sorted than is typical for inland dunes, an observation consistent with the proximal source area and stratigraphic evidence for negligible dune migration. The consistently finer grain size and better sorting of sand on the leeward face of the dune confirms the field evidence for flow separation and suspended sediment transport over a stationary dune crest.

Using the empirical relationship between grain size and mode of transport established by Bagnold (1941), we can estimate the proportion of the grain population moved by traction, saltation, and suspension. Using the threshold grain diameters of 0.350 mm (1.5 phi) and 0.088 mm (3.5 phi) for motion by creep and motion by saltation, respectively, we estimate that between 60% and 80% of the mass of the Longmeadow dune arrived as saltation load, with <10% being carried to the site in suspension. The more poorly sorted character of the granule-rich zone at the base of the transition zone (unit 2a), its coarser texture, and its large variability in texture are consistent with its deflationary origin. The similarity in texture between the eolian mantle and the underlying dune demonstrates that both accumulated primarily as saltation load. The limited amount of silt in the eolian mantle, its upward textural uniformity, and the absence of sand-silt intergrades in other sections suggest that its deposition ceased abruptly.

Interpretation. The primary bed forms preserved in exposures through the dune permit a straightforward interpretation of dune growth. The initial condition, indicated by the dominance of rippleform strata inclined in various directions, was that of a subhorizontal sand sheet accreting on a dry, unvegetated surface and with an abundant sand supply. Evidence for the infiltration of rainfall, for surface runoff, and for seasonal freezing and thawing indicates that surface moisture was present, perhaps for much of each year. Continued upward growth created an elongate mound of sand with a well-defined crest line but lacking a slip face (dome dune; ribar). When the dune reached a threshold height of about 10 m, a zone of flow separation developed, creating a southwest-directed slip face. Soon after its formation, the slip face was buried by sand carried in suspension over the dune brink deposits. The upward decrease in variability of paleowind direction (Fig. 10) and the absence of evidence for net migration indicate that the dune formed rapidly in place. Finally, the original dune form was modified by oblique and secondary winds, especially those with a westerly component of flow.

Stabilization of the dune was initially a gradual process. The granule-rich lag horizon at the base of the transitional zone and the truncation of bed forms indicates that the present dune morphology is erosional. The subhorizontal banding in the lower transition zone reflects saltation and intermittent periods of net accretion on a partially vegetated surface, an interpretation consistent with the dispersed (bioturbated) characteristic of the underlying lag horizon, and by the increasing abundance of rootlet mottling upward. Truncation structures above the incipient paleosol indicate intermittent deposition within blowouts on a largely stabilized surface prior to growth of a pine forest.

The rapid transition to the massive eolian mantle indicates an abrupt change in dune growth. Although winds remained strong enough to transport sand by saltation, conditions associated with uniform traction and ripple migration could not occur. An alternative mode of transport involving modified saltation and short-term suspension is likely to have developed in response to localized irregularities caused by vegetation. We cannot determine to what extent the massive character of the deposit was caused by the cooccurrence of bioturbation and sand deposition.

Over most of the dune, the A horizon of the Holocene soil is buried by as much as 50 cm of unweathered sand, in which weak ochric horizons have recently developed. Burial of the Holocene soil indicates that local deflation and sand transport accompanied historic disturbances associated with colonial forestry and farming in this area.

DISCUSSION

Moist Periglacial Regime (ca. 21 ka to ca. 12.7 ka)

The absence of transverse dunes and barchanoid ridges on older deglacial deposits indicates that significant accretion of dunes did not occur early in the deglacial hemicycle. Additionally, the absence of primary eolian cross-bedding in the nearly ubiquitous sand sheets suggests that summer conditions remained moist. This interpretation is consistent with the apparent delay in the construction of the Longmeadow dune. Although the lake deposits on which it was later built were exposed as early as 16 ka, and although lacustrine spits that formed by strong northeasterly winds were developing during this interval, bedded sand sheets and dunes apparently did not develop until after about 12–13 ka.

A climatic restriction on dune growth earlier in the deglacial phase is supported by the dominance of sedge and willow (rather than grasses and forbs) in the tundra pollen zone and with the results of paleoclimatic simulations requiring heavy snowfalls and increased soil moisture in the periglacial zone. This interpretation is further supported by the conclusions of Gustavson and Boothroyd (1987), who compared New England's early deglacial climate to that of the Malaspina Glacier, Alaska, where permafrost is absent, where annual precipitation exceeds 1 m/yr, and where eolian deposits form a negligible component of the sedimentary system. The climatic mechanism responsible for maintaining summer wetness in New England during a time of strong katabatic outflow cannot be extracted from our geological evidence. However, we speculate that enhanced summer wetness was caused by a combination of cooler summer temperatures and heavier summer rainfalls from cyclonic storms, both of which would have been influenced by the local presence of the Laurentide ice sheet.

Dry Periglacial Regime (ca. 12.7 ka to ca. 12.4 ka)

The massive size of the Longmeadow dune, negligible migration, unimodal paleowinds, and the absence of secondary structures over most of the dune indicate that it accreted rapidly, perhaps only a few centuries before lake drainage. Such rapid accretion on a surface that had been previously exposed for at least a millennium indicates that dune growth was initiated by an abrupt shift toward drier summer conditions, a transition that apparently took place more quickly than vegetation could respond as a stabilizing influence.

Within the limits of dating uncertainties, the change toward rapid dune growth coincides with the beginning of the Bolling/Allerod interval at about 12.7 ka (Broecker, 1992), a transition that probably took place in a few decades or less. Dune growth also coincided with a period of strong ice-sheet melting (Fairbanks, 1989) and an abrupt shift toward warmer sea-surface temperatures in the North Atlantic (Lehman and Keigwin, 1992). A comparable warmth-induced development of late-glacial dunes also occurred in northern Europe and in interior Alaska (Lea and Waythomas, 1990; Koster and Dijkmans, 1988).

Primary and secondary sedimentary structures within the Longmeadow dune indicate sand transport by gusty (>6 m/s), presumably anticylonic winds under dry summer conditions at a distance of at least 150 km from the contemporary ice margin. The consistent southwesterly dip of grain-flow strata in the Longmeadow dune, the facies most useful for reconstructing paleowinds (Sweet, 1992), in conjunction with the extensive ripple-form strata, demonstrate that the sand-moving winds were northeasterly and nearly unimodal (26.5 ± 4.2). Such a northeasterly regime confirms the katabatic outflow of descending air from the Laurentide ice sheet, because neither the valley floor topography nor an unmodified gravity flow could have generated the easterly component.

Notably absent from the Longmeadow dune are structures that indicate the codeposition of blowing snow and sand. Apparently, all dune movement occurred after complete snowmelt on the south-facing leeward slope. Also absent are significant cryogenic disturbances, features that are ubiquitous in eolian facies where permafrost is known to have been present (Lea, 1990). The absence of ice-wedges, ground cracks, and cryoturbated strata across nearly 200 m of continuous exposure strongly suggests that permafrost was absent during dune accretion, at least on this well-drained site. The apparent absence of permafrost here bears no relationship to the likely development of permafrost earlier in the deglacial hemicycle (Stone and Ashley, 1992).

The abrupt increase in spruce pollen across the deglaciated northeast dated between about 12.3 and 12.5 ka (Davis, 1983) and the nearly identical ages for the first appearance of AMS-dated spruce macrofossils in eastern New York (Peteet et al., 1990) appear to coincide with the drainage of Glacial Lake Hitchcock and to postdate growth of the Longmeadow dune by only a few centuries. Although speculative, the relative sequence of these events suggests that the abrupt amelioration at the beginning of the Bolling/Allerod interval contributed to a reduction in the height and extent of the Laurentide ice sheet below the threshold required to generate a summer anticyclone, allowing the penetration of warm westerly airflow and the northeasterly transport of arboreal pollen. The abrupt increase in coniferous pollen could thus be an indirect, delayed response caused by a change in prevailing winds during the pollen season. Additionally, the drainage of Glacial Lake Hitchcock could also be an indirect response to the same climate warming; sharply increased melt rates for buried ice in the Hitchcock dam may have contributed to its failure.

The Northwesterly Regime (ca. 12.4–11 ka)

The consistent northwesterly trend of parabolic dunes on the floor of Glacial Lake Hitchcock indicates the breakdown of the glacial anticyclone by the time of lake drainage (12.4 ka?) and its replacement by a west-northwesterly prevailing regime similar to that of today (Wells, 1983). Such localized pulses of eolian activity caused by the sudden availability of sand took place when summer conditions were dry enough to permit sand movement (Webb et al., 1993). The westerly regime coincides with the transitional zone in the stratigraphy at Longmeadow. We speculate that the sudden upwind exposure of an unvegetated lowland led to a localized increase in the intensity and frequency of near-surface, deflationary winds. Such an increase would have gradually diminished as vegetation reclaimed the abandoned lake floor and the interdune lowlands. Dune stabilization apparently culminated with paleosol development and the growth of pine forest.

The interpretations above are most consistent with the drainage of Glacial Lake Hitchcock in Connecticut at about 12.4 ka. Although Stone and Ashley (1992) suggested an earlier date for lake drainage (13.5–13.7 ka), their interpretation cannot easily explain the absence of transverse dunes on the lake bottom deposits and the absence of evidence for a switchover in wind regime.[3]

The Younger Dryas Regime (ca. 11–10 ka)

Renewed accumulation of eolian sand occurred at Longmeadow soon after 11.5 ka. The increased silt content, pervasive oxidation, and the complete absence of primary bedding indicate that sand accumulation occurred under conditions that were moister than those associated with the underlying deposits, although the sediment texture indicates that wind intensity remained strong. The eolian mantle at Longmeadow may thus be sedimentologically equivalent to undated parts of the eolian mantle elsewhere, where summer moisture inhibited bed-form development.

This abrupt transition to wetter conditions coincides with the Younger Dryas interval, an abrupt climatic reversal to near-glacial conditions at high-latitudes, dated at between 11 ka and 10 ka. Peteet et al. (1990) confirmed a Younger Dryas pollen signal for New England, in which the reversal to more boreal taxa between 11 ka and 10 ka is interpreted as a return to cooler conditions. At Longmeadow, the abrupt change from finely bedded dune sand to massive, oxidized, silty sand sometime after 11.5 ka (13 400 calendar yr) suggests that the Younger Dryas was also associated with enhanced soil moisture, and with gusty winds capable of sand traction on a vegetated surface. The absence of simultaneous eolian deposition in the "pingo-scars" indicates that the availability of sand was restricted to pre-existing dunes. Sand transport during the Younger Dryas reversal may have ended rapidly, a speculation consistent with the downwind inception of posteolian soils at or before 10.6 ka (Thorson and McBride, 1988).

The Holocene Regime (ca. 10.5 ka to Present)

Reactivation of dunes in the Longmeadow area apparently did not occur at any time during the Holocene. In contrast, dune reactivation at Longmeadow and elsewhere in New England was associated with historic agricultural and forestry practices.

ACKNOWLEDGMENTS

GSA reviewers Julie Brigham-Grette, William McCoy, and Janet Stone provided many helpful suggestions. Robert Oldale encouraged us to proceed with this review, despite the qualitative and spotty nature of the previous literature. Partial support for this project was provided by a grant from the Connecticut Watershed Council. The accelerator date and the charcoal identification were provided by Lucinda McWeeney.

REFERENCES CITED

Ahlbrandt, T. S., 1979, Textural parameters of eolian deposits, *in* McKee, E. D., eds., A study of global sand seas: U.S. Geological Survey Professional Paper 1052, p. 21–51.

Ahlbrandt, T. S., and Andrews, S., 1978, Distinctive sedimentary features of cold-climate eolian deposits, North Park, Colorado: Palaeogeography, Palaeoclimatology, and Palaeoecology, v. 25, p. 327–351.

[3]The older estimate for lake drainage is a reversal from the previous conclusion of Lewis and Radway-Stone (1991), which placed the date for lake drainage at about 12.4 ka, based on basal dates from the peat-filled depressions (Pingos?).

Antevs, E., 1922, The recession of the last ice sheet in New England: American Geographical Society Research Series Serial No. 17, 292 p.
Bagnold, R. A., 1941, The physics of blown sand and desert dunes: London, Metheuen, 256 p.
Barnosky, C. W., Anderson, P. M., and Bartlein, P. J., 1987, The northwestern U.S. during deglaciation; Vegetational history and paleoclimatic implications, *in* Ruddiman, W. F., and Wright, H. E., Jr., eds., North America and adjacent oceans during the last deglaciation: Boulder, Colorado, Geological Society of America, Geology of North America, v. K-3, p. 289–321.
Black, R. F., 1983, Pseudo–ice wedge casts of Connecticut, northeastern United States: Quaternary Research, v. 20, p. 74–89.
Broecker, W. S., 1992, Defining the boundaries of the late-glacial isotope episodes: Quaternary Research, v. 38, p. 135–138.
Bryan, K., 1931, Wind-worn stones or ventifacts—A discussion and bibliography: National Research Council Report to the Committee on Sedimentation, 1929–1930, p. 29–50.
Clark, P. U., 1992, Surface form of the southern Laurentide ice sheet and its implications to ice-sheet dynamics: Geological Society of America Bulletin, v. 104, p. 595–605.
Clebnik, S. M., 1980, Surficial geology of the Willimantic quadrangle: Connecticut Geological and Natural History Survey Quadrangle Report QR-39, 45 p.
COHMAP, 1988, Climatic changes of the last 18,000 years; Observations and model simulations: Science, v. 241, p. 1043–1052.
Colton, R. B., 1960, Surficial geologic map of the Windsor Locks quadrangle, Connecticut: U.S. Geological Survey Geologic Quadrangle Map GQ-137, scale 1:24 000.
Colton, R. B., 1962, Lacustrine features in north-central Connecticut: Geological Society of America Special Paper 68, p. 155.
Colton, R. B., 1965a, Surficial geological map of the Broad Brook quadrangle, Hartford and Tolland counties, Connecticut: U.S. Geological Survey Geologic Quadrangle Map GQ-223, scale 1:24 000.
Colton, R. B., 1965b, Surficial geologic map of the Manchester quadrangle, Connecticut: U.S. Geological Survey Geologic Quadrangle Map GQ-433, scale 1:24 000.
Cushman, R. V., 1963, Geology of the Hartford North quadrangle, Connecticut: U.S. Geological Survey Geologic Quadrangle Map GQ-223, scale 1:24 000.
David, P. P., 1981, Stabilized dune ridges in northern Saskatchewan: Canadian Journal of Earth Sciences, v. 18, p. 286–310.
David, P. P., 1989, Eolian processes, *in* Fulton, R. J., ed., Quaternary geology of Canada and Greenland: Ottawa, Geological Survey of Canada, Geology of Canada series, v. 1 (Geological Society of America, Geology of North America, v. K-1), p. 620–623.
Davis, M. B., 1983, Holocene vegetational history of the eastern United States, *in* Wright, H. E., Jr., ed., Late Quaternary environments of the United States, Volume 2: The Holocene: Minneapolis, University of Minnesota Press, p. 161–181.
Denny, C. S., 1936, Periglacial phenomena in southern Connecticut: American Journal of Science, v. 32, p. 322–342.
Dineen, R. J., DeSimone, D. J., and Hanson, E. L., 1988, Glacial Lake Albany and its successors in the Hudson Lowlands, *in* Brigham-Grette, J., ed., American Quaternary Association, 1988 Meeting, Field Trip Guidebook: Amherst, Massachusetts, American Quaternary Association, p. 15–16.
Fairbanks, R. G., 1989, A 17,000-year glacio-eustatic sea level record: Influence of glacial melting rates on the Younger Dryas event and deep-ocean circulation: Nature, v. 342, p. 637–642.
Filion, L., 1987, Holocene development of parabolic dunes in the central St. Lawrence Lowland, Quebec: Quaternary Research, v. 28, p. 196–209.
Flint, R. F., 1930, The glacial geology of Connecticut: Connecticut Geological and Natural History Survey Bulletin 47, 294 p.
Folk, R. L., 1968, Petrology of sedimentary rocks: Austin, Texas, Hemphill Press, 170 p.
Forman, S. L., Goetz, A.F.H., and Yuhas, R. H., 1992, Large-scale stabilized dunes on the high plains of Colorado: Understanding the landscape response to Holocene climates with the aid of images from space: Geology, v. 20, p. 145–148.
Freyberger, S. G., 1979, Dune forms and wind regime, *in* McKee, E. D., ed., A study of global sand seas: U.S. Geological Survey Professional Paper 1052, p. 137–171.
Freyberger, S. G., Ahlbrant, T. S., and Andrews, S., 1979, Origin, sedimentary features, and significance of low-angle sand sheet deposits, Great Sand Dunes National Monument Vicinity, Colorado: Journal of Sedimentary Petrology, v. 49, p. 733–746.
Fuller, M. L., 1914, The geology of Long Island, New York: U.S. Geological Survey Professional Paper 82, 231 p.
Gustavson, T. C., and Boothroyd, J. C., 1987, A depositional model for outwash, sediment sources, and hydrologic characteristics, Malaspina Glacier, Alaska: A modern analog of the southeastern margin of the Laurentide ice sheet: Geological Society of America Bulletin, v. 99, p. 292–302.
Hartshorn, J. H., 1962, Late-glacial eolian activity in Massachusetts: Geological Society of America Special Paper 68, p. 194.
Hartshorn, J. H., 1967, Geology of the Taunton quadrangle, Bristol and Plymouth Counties, Massachusetts: U.S. Geological Survey Bulletin 1163-D, p. D1–D67.
Hartshorn, J. H., and Colton, R. B., 1971, Surficial geologic map of the West Springfield quadrangle, Massachusetts and Connecticut: U.S. Geological Survey Geologic Quadrangle Map GQ-892, scale 1:24 000.
Hartshorn, J. H., and Koteff, C., 1967, Geologic map of the Springfield South quadrangle, Massachusetts-Connecticut: U.S. Geological Survey Geologic Quadrangle Map GQ-678, scale 1:24 000.
Howard, J. D., 1975, The sedimentological significance of trace fossils, *in* Frey, R. W., ed., The study of trace fossils: New York, Springer-Verlag, p. 131–146.
Hunter, R. E., 1977, Basic types of stratification in small eolian dunes: Sedimentology, v. 24, p. 361–387.
Jahns, R. H., and Willard, M. E., 1942, Late Pleistocene and recent deposits in the Connecticut Valley, Massachusetts: American Journal of Science, v. 240, p. 161–191.
Johnson, D. L., 1990, Biomantle evolution and the redistribution of earth materials and artifacts: Soil Science, v. 149, p. 84–101.
Karrow, P. F., and Occhietti, S., 1989, Quaternary geology of the St. Lawrence Lowlands of Canada, *in* Fulton, R. J., ed., Quaternary geology of Canada and Greenland: Ottawa, Geological Survey of Canada, Geology of Canada series, v. 1 (Geological Society of America, Geology of North America, v. K-1), p. 321–389.
Kocurek, G., and Fielder, G., 1982, Adhesion structures: Journal of Sedimentary Petrology, v. 52, p. 1229–1241.
Koster, E. A., and Dijkmans, J. W. A., 1988, Niveo-aeolian deposits and denivation forms, with special reference to the Great Kobuk sand dunes, northwestern Alaska: Earth Surface Processes and Landforms, v. 13, p. 153–170.
Koteff, C., Stone, J. R., Larsen, F. D., Ashley, G. M., Boothroyd, J. C., and Dincauze, D. F., 1988, Glacial Lake Hitchcock, postglacial uplift and post-lake archaeology, *in* Brigham-Grette, J., ed., American Quaternary Association, 1988 Meeting, Field Trip Guidebook: Amherst, Massachusetts, American Quaternary Association, p. 169–208.
Kutzbach, J. E., and Guetter, P. J., 1986, The influence of changing orbital patterns and surface boundary conditions on climate simulations for the past 18,000 years: Journal of Atmospheric Science, v. 43, p. 1726–1759.
Kutzbach, J. E., and Wright, H. E., Jr., 1985, Simulation of the climate of 18,000 years B.P.: Results for the North American/North Atlantic/European sector and comparison with the geologic record of North America: Quaternary Science Reviews, v. 4, p. 147–187.
Lea, P. D., 1990, Pleistocene periglacial eolian deposits in southwestern Alaska; Sedimentary facies and depositional processes: Journal of Sedimentary Petrology, v. 60, p. 582–591.
Lea, P. D., and Waythomas, C. F., 1990, Late-Pleistocene eolian sand sheets in Alaska: Quaternary Research, v. 34, p. 269–281.
Lehman, S. J., and Keigwin, L. D., 1992, Sudden changes in north Atlantic circulation during the last deglaciation: Nature, v. 356, p. 757–762.
Lewis, R. S., and Radway-Stone, J., 1991, Late Quaternary stratigraphy and depositional history of the Long Island Sound Basin: Connecticut and New York: Journal of Coastal Research, v. 11, p. 1–23.
Lougee, R. J., 1939, Geology of the Connecticut watershed: New Hampshire Fish and Game Department, Biological Survey of the Connecticut Watershed Survey Report 4, p. 131–149.
Manabe, S., and Broccoli, A. J., 1985, The influence of continental ice sheets on the climate of an ice age: Journal of Geophysical Research, v. 90, p. 2167–2190.
Mather, K. F., Goldthwait, R. P., and Thiesmeyer, L. R., 1942, Pleistocene geology of western Cape Cod, Massachusetts: Geological Society of America Bulletin, v. 53, p. 1127–1174.
Mayewski, P. A., Meeker, L. D., Whitlow, S., Twickler, M. S., Morrison, M. C., Alley, R. B., Bloomfield, P., and Taylor, K., 1993, The atmosphere during the Younger Dryas: Science, v. 261, p. 195–200.
McKee, E. D., 1979, Sedimentary structures in dunes, *in* McKee, E. D., ed., A study of global sand seas: U.S. Geological Survey Professional Paper 1052, p. 87–134.
McKee, E. D., and Bigarella, J. J., 1972, Deformational structures in Brazilian coastal dunes: Journal of Sedimentary Petrology, v. 42, p. 670–681.
McKee, E. D., and Weir, G. W., 1953, Terminology for stratification and cross-stratification in sedimentary rocks: Geological Society of America Bulletin, v. 64, p. 381–390.
McKeon, J. B., 1989, Late glacial dunes, ventifacts and wind direction in west-central Maine, *in* Tucker, R. D., and Marvinney, R. G., eds., Studies in Maine geology, Volume 6: Quaternary geology: Augusta, Maine Geological Survey, p. 89–101.
Mickelson, D. M., Clayton, L., Fullerton, D. S., and Borns, H. W., Jr., 1983, The late Wisconsin glacial record in the United States, *in* Wright, H. E., Jr., ed., Late Quaternary environments of the United States, Volume 1: The late Pleistocene: Minneapolis, University of Minnesota Press, p. 3–37.
Oldale, R. N., and O'Hara, C. J., 1984, Glaciotectonic origin of the Massachusetts coastal end moraines and a fluctuating late Wisconsinan ice margin: Geological Society of America Bulletin, v. 95, p. 61–74.
Peteet, D. M., Vogel, J. S., Nelson, D. E., Southon, J. R., Nickmann, R. J., and Heusser, L. E., 1990, Younger Dryas climatic reversal in northeastern USA? AMS ages for an old problem: Quaternary Research, v. 33, p. 219–230.
Ridge, J. C., and Larsen, F. D., 1990, Re-evaluation of Antev's New England varve chronology and new radiocarbon-dates of sediments from Glacial Lake Hitchcock: Geological Society of America Bulletin, v. 102, p. 889–899.
Rind, D., 1987, Components of the ice age circulation: Journal of Geophysical Research, v. 92, p. 4241–4281.
Schafer, J. P., and Hartshorn, J. H., 1965, The Quaternary of New England, *in* Wright, H. E., Jr., and Frey, D. G., eds., The Quaternary of the United States: Princeton, New Jersey, Princeton University Press, p. 113–128.
Schile, C. A., 1991, Stationary dune growth in the Connecticut River valley (Springfield South quadrangle) [M.S. thesis]: Storrs, University of Connecticut, 82 p.
Schwan, J., 1988, The structure and genesis of Weichselian to early Holocene aeolian sand sheets in western Europe: Sedimentary Geology, v. 55, p. 197–232.
Sharp, R. P., 1980, Wind-driven sand in Coachella Valley, California; Further data: Geological Society of America Bulletin, v. 91, p. 724–730.
Smith, H. T. U., and Fraser, H. J., 1935, Loess in the vicinity of Boston, Massachusetts: American Journal of Science, ser. 5, v. 30, no. 175, p. 16–32.
Stone, B. D., and Borns, H. W., 1986, Pleistocene glacial and interglacial stratigraphy of New England, Long Island and adjacent Georges Bank and Gulf of Maine, *in* Sibrava, V., Bowen, D. Q., and Richmond, G. M., eds., Quaternary glaciations in the Northern Hemisphere: Quaternary Science Reviews, v. 5, p. 39–52.
Stone, J. R., and Ashley, G. M., 1992, Ice-wedge casts, pingo scars, and the drainage of Glacial Lake Hitchcock, *in* Robinson, P., and Brady, J. B., eds., New England Intercollegiate Geological Conference, 84th Annual Meeting, Guidebook for fieldtrips in the Connecticut Valley region of Massachusetts and adjacent states: University of Massachusetts Department of Geology and Geography Contribution No. 66, p. 305–331.
Stone, J. R., Schafer, J. P., London, E. H., and Thompson, W. B., 1992, Surficial materials map of Connecticut: U.S. Geological Survey Special Geologic Map, scale 1:125 000.
Stuiver, M., Braziunas, T. F., Becker, B., and Kromer, B., 1991, Climatic, solar, oceanic, and geomagnetic influences on late-glacial and Holocene atmospheric $^{14}C/^{12}C$ change: Quaternary Research, v. 25, p. 1–24.
Sweet, M. L., 1992, Lee-face airflow, surface processes, and stratification types: Their significance for refining the use of eolian cross-strata as paleocurrent indicators: Geological Society of America Bulletin, v. 104, p. 1528–1538.
Thorson, R. M., and McBride, K., 1988, The Bolton Spring site, Connecticut; Early Holocene human occupation and environmental changes in southern New England: Geoarchaeology, v. 3, p. 221–234.
Thorson, R. M., and Webb, R. S., 1991, Postglacial history of a cedar swamp in southeastern Connecticut: Journal of Paleolimnology, v. 6, p. 17–35.
Webb, T., III, Bartlein, P. J., Harrison, S. P., and Anderson, K. H., 1993, Vegetation, lake levels, and climate in eastern North America for the past 18,000 years, *in* Wright, H. E., Jr., Kutzbach, J. E., Webb, T., III, Ruddiman, W. F., Street-Perrott, F. A., and Bartlein, P. J., eds., Global climates since the last glacial maximum: Minneapolis, University of Minnesota Press, p. 415–467.
Webb, T., III, Bartlein, P. J., and Kutzbach, J. E., 1987, Climatic change in eastern North America during the past 18,000 years; Comparisons of pollen data with model results, *in* Ruddiman, W. F., and Wright, H. E., Jr., eds., North America and adjacent oceans during the last deglaciation: Boulder, Colorado, Geological Society of America, Geology of North America, v. K-3, p. 447–462.
Wells, G. L., 1983, Late-glacial circulation over North America revealed by aeolian patterns, *in* Street-Perrott, A., Beran, M., and Ratcliff, R., eds., Variations in the global water budget: Boston, Reidel, p. 317–330.

Manuscript Received by the Society January 18, 1994
Revised Manuscript Received September 26, 1994
Manuscript Accepted November 15, 1994

Printed in U.S.A.