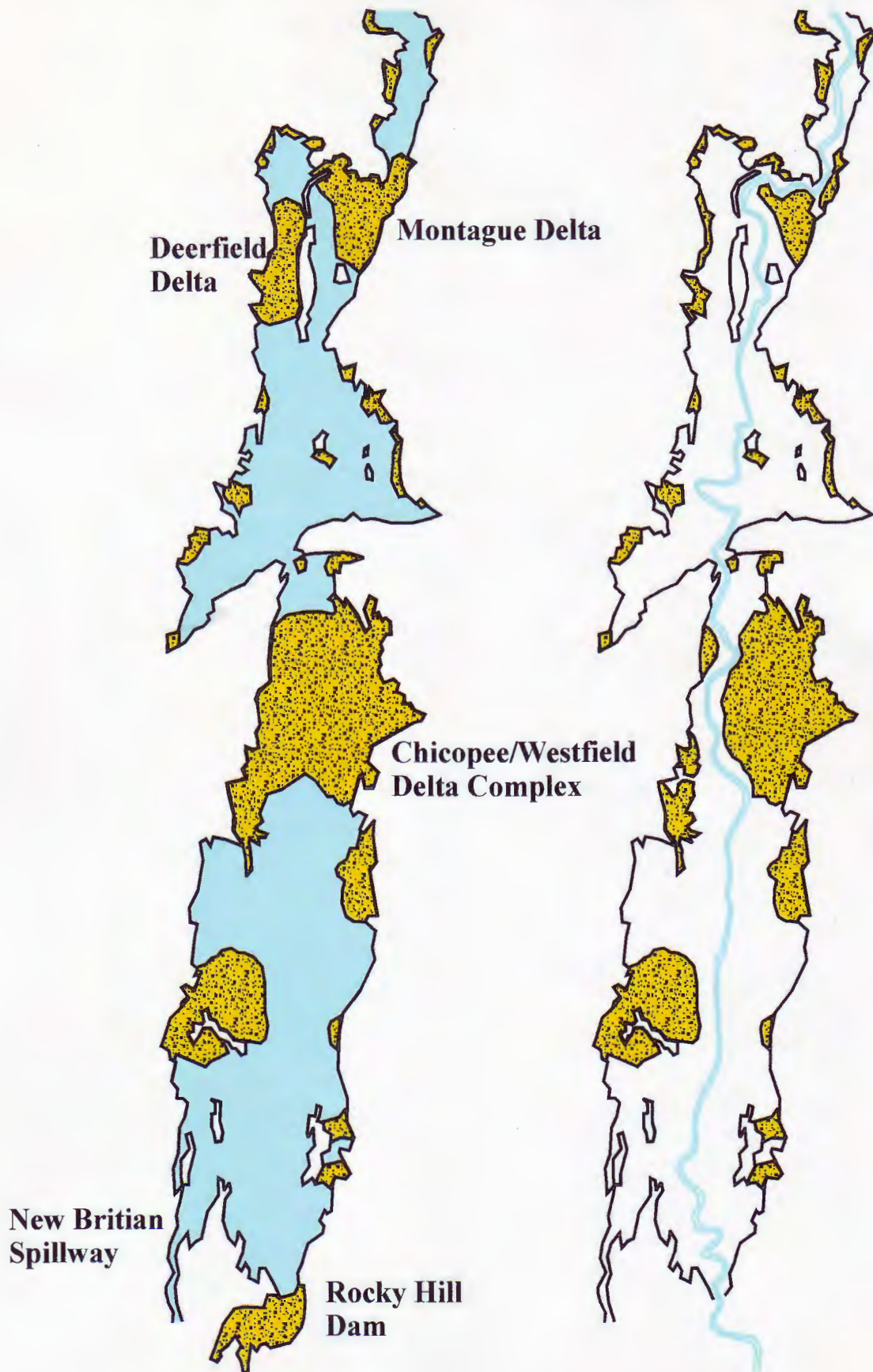


North Eastern FOP 2000

A New Drainage History for Glacial Lake Hitchcock



A New Drainage History for Glacial Lake Hitchcock: Varves, Landforms, and Stratigraphy

63rd Annual Reunion

North Eastern Friends of the Pleistocene Field Conference

June 2-4, 2000,

The Inn at Northampton, Massachusetts

The trip leaders

Julie Brigham-Grette (UMass-Amherst),
Tammy Rittenour (University of Nebraska),
Janet Stone (USGS – Water Resources Division),
Jack Ridge (Tufts University),
Al Werner, Laura Levy (Mt Holyoke College)
Dena Dincauze, Kit Curran (UMass-Amherst)
Ed Klekowski, (UMass-Amherst),
and Richard Little (Greenfield Community College).

Department of Geosciences Contribution No. 73

University of Massachusetts

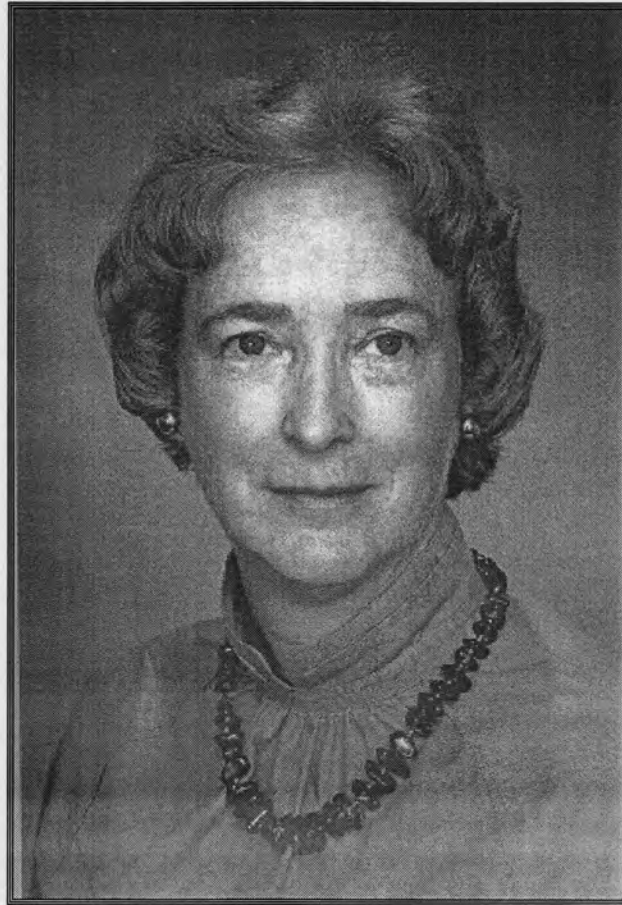
Amherst, MA 01003-5820

<http://www.geo.umass.edu>

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*****Dedication*****



Professor Dena Dincauze has worked on the archeology of New England nearly all of her professional life. Her name is synonymous with what we know regarding the late glacial/Holocene history of human occupations throughout the region. After working at the Peabody Museum and Department of Anthropology at Harvard and the Anthropology Department at SUNY-Buffalo, Dena came to UMass in the Fall of 1973 to start a 27 year career at UMass-Amherst training and engaging both undergraduate and graduate students in archeology and geoarchaeology. She also assisted the science community at large by serving, for example, as President and President-elect of the Society of Professional Archeologists (1983-1985), as President and President-elect of the Society of American Archaeology (1985-1989), and as editor of *American Antiquity* from 1980-1984, among many other outstanding accomplishments. Her latest book, *Environmental Archaeology: Principles and Practice*, (Cambridge University Press, 350 pgs.), will be out in August, 2000.

Upon the eve of Dena's retirement in December, 2000, we dedicate this FOP volume to her for her friendship and endless dedication to all the dimensions of teaching, research and service to which we can all aspire.

Schedule of FOP Events June 2-4, 2000

Friday June 2

6-10 PM Welcoming Icebreaker and Registration in the Hampshire Room

The Inn at Northampton (**Trip Headquarters**)

One Atwood Dr. (Interstate 91, Exit 18)

Northampton, MA 01060

Cash Bar available and optional pickup of registration materials

Saturday June 3

7:00-8:00 Breakfast, pick up registration materials. At the Inn, Coffee is served at 7 AM and buffet breakfast is available at 7:30.

8:15 AM Field Trip leaves from parking lot at The Inn at Northampton.
All participants are expected to ride on the buses provided for the trip (includes restroom facilities). Lots of parking provided for cars at The Inn at Northampton for those driving from elsewhere.

~5:30 PM Return to the hotel

6:30 PM Cash Bar

7:30 PM Buffet dinner, cash bar, and small program

Sunday June 4

7:00-8:00 Breakfast (. At the Inn, Coffee is served at 7 AM and buffet breakfast is available at 7:30 AM.

8:15 AM Field Trip leaves from parking lot at
The Inn at Northampton All participants expected on the buses provided for the trip (includes restroom facilities)

1:00 PM Approximate return to Northampton area
(trip continues in the Amherst/Northampton area after collecting vehicles after lunch)

3:30 PM Approximate end of the field excursion.
Return to the hotel

Trip Announcement

**63rd North East Friends of the Pleistocene
June 2-4, 2000, in the Connecticut Valley
A New Drainage History for Glacial Lake Hitchcock:
Varves, Landforms, and Stratigraphy**

This field conference will focus on the classic Late Pleistocene drainage history of Glacial Lake Hitchcock, the large pro-glacial lake system that occupied the length of the Connecticut Valley in central New England. Since 1987 (50th NE FOP), new research by a number of workers has shown that the drainage history was extremely complex and intimately linked to the retreat of the Laurentide Ice Sheet. We will examine evidence that the lake, drained sequentially as a series of subbasins -- a process that likely took thousands of years -- associated with the downcutting of the early Connecticut River through the ancient lake floor as well as through a complex system of ice-contact and meteoric deltas throughout the valley. We will examine the stratigraphy and morphology of glacial, glaciofluvial and lacustrine (varves!) deposits associated with deglaciation along with post-Hitchcock dune complexes. New geochronological control provided by ^{14}C and optically-stimulated luminescence ages shed new light on both the timing of events and gaps in our understanding of the drainage sequence coupled with regional glacioisostatic rebound. The trip headquarters will be at The Inn at Northampton, MA, same site as the 50th FOP.

The trip leaders include Janet Stone (USGS), Jack Ridge (Tufts), Tammy Rittenour (Univ. Nebraska), Julie Brigham-Grette (UMass), and Al Werner (Mt Holyoke College) with cameo appearances by Dena Dincauze (UMass), Ed Klekowski (UMass), and Richard Little (Greenfield CC).

Registration Fee: \$65 (includes two lunches, snacks, comfortable buses, guidebook and theme T-shirt): \$45 for Students (for the same) DUE APRIL 30th.

Buffet Dinner: \$35
(all you can eat)
REGISTER BY MAY 1ST PLEASE

Make checks out to
"Geosciences Fund - 6-30220"

Send registration and dinner fee to
Julie BG-FOP
Department of Geosciences
University of Massachusetts
Amherst, MA 01003
Phone: 413-545-4840 FAX: 413-545-1200

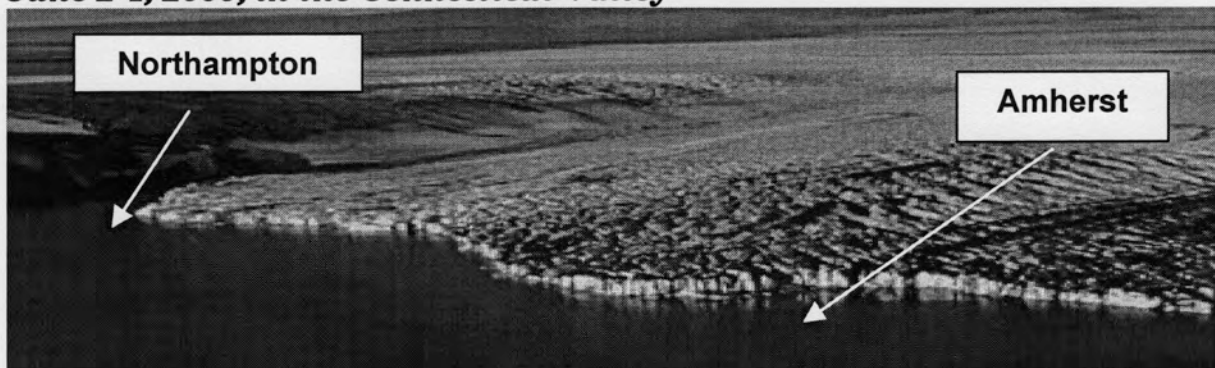
Below is a list of area hotels and *price estimates* for June 2-3. You are **strongly encouraged** to make your lodging arrangements **ASAP** due to other university/college activities in the valley the same weekend. Other area hotels and restaurants are listed at <http://virtual-valley.com>.

1. **The Inn at Northampton (**Trip Headquarters**)**
One Atwood Dr. (I/91, Exit 18)
Northampton, MA 01060
413-586-1211
800-582-2929
email: innoho@javanet.com, <http://virtual-valley.com/innatoho>
\$119.00 + tax (1-4 people)
(MENTION "FOP" TO GET THIS SPECIAL RATE UNTIL APRIL 30TH; AFTER THAT NORMAL RATES APPLY)
2. **The Valley Inn- Best Western**
117 Conz St., Northampton
413-586-1500
800-941-3066
\$99.00 + tax (2 people)
3. **Howard Johnsons**
401 Russell St. /Rte. 9
Amherst/Hadley
413-586-0114
<http://virtual-valley.com/hojo>
\$115.00 + tax (2 people)
4. **Motel 6**
State Road / Deerfield
2 exits north of The Inn at Northampton
413-665-7161
nationwide reservations 800-466-8356
\$51.99 + tax (2 people)
5. **Norwottuck Inn**
208 Russell St. (Rte. 9)
Hadley, MA
413-587-9866
\$85.00 + tax (2 people)
6. **Country Belle Motel**
Rte. 9 / Russell St., Hadley
413-586-0715
\$75.00 + tax (2 people)
7. **Econo Lodge**
237 Russell St. (Rte. 9)
Hadley, MA
413-584-9816
\$75.00 + tax for 1 Queen or King
\$79.00 + tax for 2 double

Registration Form.

63rd Friends of the Pleistocene

June 2-4, 2000, in the Connecticut Valley



**A NEW DRAINAGE HISTORY FOR
GLACIAL LAKE HITCHCOCK:
VARVES, LANDFORMS, AND STRATIGRAPHY**

REGISTRATION FORM ----- MUST REGISTER BY MAY 1ST

NAME _____

ADDRESS _____

PHONE: _____; FAX: _____

EMAIL: _____

Fees: (X those that apply)

_____ \$65 (includes two lunches, snacks, comfortable buses, guidebook and theme T-shirt):

_____ \$45 for Students (for the same)

_____ Buffet Dinner: \$35 (all you can eat)

_____ Total Payment

Make checks out to

"Geosciences Fund - 6-30220"

FAX: 413-545-1200
Phone: 413-545-4840

Send registration and dinner fee to

Julie BG-FOP
Department of Geosciences
University of Massachusetts
Amherst, MA 01003

*****Notes*****

AN INTRODUCTION TO THE NE FOP 2000

Julie Brigham-Grette, Trip Leader

It is the tradition of every FOP trip to provide the faithful followers of this annual field conference with a stimulating look at new research and field data regarding the regional Quaternary history of New England. This year the trip revisits the history of Glacial Lake Hitchcock with a focus on new evidence for the timing of deglaciation, processes influencing lake sedimentation, and factors influencing the drainage sequence and style of Lake Hitchcock. In this context we will also reconsider how the drainage history may be related to the rate and timing of glacio-isostatic rebound. The fundamental challenge of this work is to evaluate changes in sub-basin varve history relative to the sequential incision and evolution of the Connecticut River. New AMS-¹⁴C age estimates will be presented to further anchor the Antev's varve chronology. These new ages and revisions to the regional varve sequence provide us with new opportunities for evaluating the varve record as a proxy for high resolution paleoclimate (including teleconnections with El Nino events) in comparison with changes in North Atlantic thermohaline circulation and rapid environmental change over the Greenland Ice Sheet (c.f., Ridge et al., 1999; Rittenour et al., 2000). We have also taken advantage of newer geochronological techniques including the optical luminescence dating of dune complexes on geomorphic surfaces of different relative age. Not to be overlooked is the work of Ed Klekowski and grad student Sean Werle discovering the underwater world of the Connecticut River related to the special habitat offered by eroding submerged varves.

This trip occurs some 12 years after the last FOP in this region which had a similar theme. Though some may have assumed we once had it all figured out based on the work presented at the 1987 FOP, it is the axiom of science that the more we learn, the more we realize just how little we know. The notion that Glacial Lake Hitchcock along its entire length drained as a result of the erosion of the dam at Rocky Hill has now given way to a new paradigm involving the northward sequential drainage of the lake as a broken series of sub-basins controlled by local dams. Yet the most robust feature of the valley history is the observation that most of the ice-contact deltas all project onto a smooth straight line ("Koteff's curve") suggesting uplift progressed without warping of the regional lake shorelines. The apparent lack of restrained rebound during deglaciation and the need to invoke delayed post-glacial rebound until glacial ice was out of the Connecticut valley is difficult given revisions in lake drainage. What have we overlooked?

I would like to take this opportunity to thank all of the trip leaders for their contributions to this guidebook and to the science presentations at the road stops. I especially want to thank Trent Hayden and Celeste Cosby, two of my current graduate students who were continuously willing to help out with the management of the registration forms, headcounts, envelope stuffing, local arrangements, and more.

Article A. RIDGE ET AL, 1999

Varve, paleomagnetic, and ^{14}C chronologies for Late Pleistocene Events in New Hampshire and Vermont (U.S.A.), *Géographie physique et Quaternaire*, v, 53, o.1, 79-106

Géographie physique et Quaternaire, 1999, vol. 53, n° 1, p. 79-106, 15 fig., 2 tabl., 1 app.

Cheers,
John Ridge

VARVE, PALEOMAGNETIC, AND ^{14}C CHRONOLOGIES FOR LATE PLEISTOCENE EVENTS IN NEW HAMPSHIRE AND VERMONT (U.S.A.)

John C. RIDGE*, Mark R. BESONEN, Marc BROCHU, Sarah L. BROWN, Jamie W. CALLAHAN, Glenn J. COOK, Robert S. NICHOLSON and Nathaniel J. TOLL, first, fifth, sixth, seventh and eighth authors : Department of Geology, Tufts University, Medford, Massachusetts 02155 U.S.A.; second author : Department of Geosciences, University of Massachusetts, Amherst, Massachusetts 01003 U.S.A.; third author : Corporate Environmental Advisors, 127 Hartwell St., West Boylston, Massachusetts 01583 U.S.A.; fourth author : Department of Geology, University of Vermont, Burlington, Vermont 05405-0122 U.S.A.

ABSTRACT A deglacial chronology for northern New England has been formulated using an atmospheric ^{14}C calibration of the New England Varve Chronology and paleomagnetic records. This ^{14}C chronology is based on ^{14}C ages from macrofossils of non-aquatic plants and is about 1500 yr younger than existing chronologies that are based primarily on ^{14}C ages of bulk organic samples. The lower and upper Connecticut Valley varve sequences of Ernst Antevs (NE varves 2701-6352 and 6601-8500) overlap (lower 6012 = upper 6601) based on their crudely matching varve records and their similar paleomagnetic records. Three ^{14}C ages at Canoe Brook, Vermont (NE varve 6150 = 12.3 ^{14}C ka) calibrate the lower Connecticut Valley sequence. New AMS and conventional ^{14}C ages on woody twigs from Newbury, Vermont calibrate the upper sequence from 11.6-10.4 ^{14}C ka (NE varves 7440-8660) and are consistent with the overlapping varve and paleomagnetic records, and the Canoe Brook ^{14}C ages. Deglaciation of the Connecticut Valley in southern Vermont began at 12.6 ^{14}C ka (15.2 cal ka) and the Littleton-Bethlehem Readvance in northern New Hampshire and Vermont reached its maximum at 11.9-11.8 ^{14}C ka (14.0-13.9 cal ka) followed by recession of ice into Québec at about 11.5 ^{14}C ka (13.4 cal ka). A lake persisted in the upper Connecticut Valley until at least 10.4 ^{14}C ka (12.3 cal ka) and may have been seen by the first humans in the area.

RÉSUMÉ Chronologie de la déglaciation au Pléistocène supérieur, au New Hampshire et au Vermont (É.-U.A.) établie à partir des varves, du paléomagnétisme et des datations au ^{14}C . La chronologie de la déglaciation du nord de la Nouvelle-Angleterre a été établie à l'aide de l'étalonnage de la New England Varve Chronology et des relevés du paléomagnétisme. Cette chronologie, fondée sur les datations au radiocarbone de macrofossiles et de plantes non aquatiques, est environ 1550 ans plus jeune que les chronologies existantes en grande partie fondées sur les datations d'échantillons organiques et de macrofossiles aquatiques et non aquatiques. Les séquences de varves des vallées inférieure et supérieure du Connecticut de Ernst Antevs (varves 2701-6352 et 6012-8500) se superposent (6012 de la vallée inférieure = 6601 de la vallée supérieure). Trois dates au ^{14}C de Canoe Brook (Vermont) (varve 6150 = 12,3 ^{14}C ka) étalonneront la séquence de la vallée inférieure du Connecticut. De nouvelles datations au ^{14}C conventionnelles et par spectrométrie de masse sur des brindilles en provenance de Newbury (Vermont) étalonneront la séquence du cours supérieur de 11,6 à 10,4 ^{14}C ka (varves 7440-8660) et concordent avec les relevés de varves chevauchantes et du paléomagnétisme et avec les datations au ^{14}C de Canoe Brook. La déglaciation de la vallée du Connecticut dans le sud du Vermont a commencé à 12,6 ^{14}C ka (15,2 ka écal.) et la récession de Littleton-Bethlehem dans le nord du New Hampshire et du Vermont était maximale à 11,9-11,8 ^{14}C ka (14,0-13,9 ka écal.) suivie du retrait de la glace au Québec vers 11,5 ^{14}C ka (13,4 ka écal.). Le lac qui a occupé la vallée supérieure du Connecticut jusqu'à environ 10,4 ^{14}C ka (12,3 ka écal.) a peut-être été aperçu par les premiers humains à fréquenter la région.

ZUSAMMENFASSUNG Warve, Paläomagnetismus und ^{14}C -Chronologien für Ereignisse im späten Pleistozän in New Hampshire und Vermont. Man hat eine Enteisungs-Chronologie für den Norden Neu-Englands formuliert, indem man eine atmosphärische ^{14}C -Eichung der New England Varve Chronology und paläomagnetische Belege benutzt hat. Diese ^{14}C -Chronologie basiert auf ^{14}C -Datierungen von Makro-Fossilien nichtaquatischer Pflanzen und ist etwa 1500 Jahre jünger als schon bestehende Chronologien, die vorwiegend auf ^{14}C -Datierungen von organischen Proben sowohl aquatischer wie nichtaquatischer Makrofossilien beruhen. Die Warven-Sequenzen vom unteren und oberen Connecticut Valley von Ernst Antevs (NO-Warven 2701-6352 und 6601-8500) überlappen sich (untere 6012 = obere 6601), sie haben grob übereinstimmende Warven-Belege und ähnliche paläomagnetische Belege. Drei ^{14}C -Datierungen bei Canoe Brook, Vermont (NO-Warve 6150 = 12.3 ^{14}C ka) kalibrieren die untere Connecticut Valley-Sequenz. Neue Massenspektrometrie- und konventionelle ^{14}C -Datierungen auf Zweigen von Newbury, Vermont, bemessen die obere Sequenz auf 11.6 bis 10.4 ^{14}C ka (NO-Warven 7440-8660) und stimmen mit der überlappenden Warve und paläomagnetischen Belegen sowie den ^{14}C -Datierungen von Canoe Brook überein. Die Enteisung des Connecticut Valley in Süd-Vermont begann um 12.6 ^{14}C ka (15.2 cal ka) und der Littleton-Bethlehem Rückvorstoß in Nord-New Hampshire und Vermont erreichte sein Maximum um 11.9-11.8 ^{14}C ka (14.0-13.9 cal ka), gefolgt von einem Eisrückzug in Richtung Québec um etwa 11.5 ^{14}C ka (13.4 cal ka). Ein See existierte im oberen Connecticut Valley bis mindestens 10.4 ^{14}C ka (12.3 cal ka) und ist vielleicht noch von den ersten Menschen in diesem Gebiet gesehen worden.

INTRODUCTION

Controversy over the style and chronology of deglaciation in northern New England has brewed for 60 years and has centered on whether the region was deglaciated by systematic ice recession from south to north (Lougee, 1935b, 1940) or by regional downwasting and stagnation (Flint, 1929, 1930, 1932, 1933; Goldthwait, 1938; Goldthwait *et al.*, 1951). At the heart of this conflict has been the analysis of varves (Antevs, 1922, 1928), which established a high-resolution chronology that was largely disregarded after the 1930's. Today disagreements center on the absolute ages of deglacial events, in particular the development of a ^{14}C chronology that will allow an accurate comparison to events in other regions of North America and the North Atlantic region.

This paper is the first comprehensive synthesis of the New England (NE) Varve Chronology in northern New England since the 1920's (Antevs, 1922, 1928). Because the varve chronology is anchored in work done at the beginning of this century we provide an historical analysis of the development of the chronology by Ernst Antevs in addition to recent developments and their implications. This historical presentation is critical because few glacial and Quaternary geologists working in New England over the last 50 years have been aware of the temporal and areal coverage of the New England Varve Chronology and the highly reproducible nature of its records in multiple drainage basins. Few studies in Quaternary geology from the early part of this century have had the scientific rigor that characterizes the varve chronology. At places in New England where varve thickness reflects the annual response of glacier melting to regional weather patterns, varve records represent a tremendous possibility for precise regional correlation. As with any Pleistocene chronology, varve records need to be tested with as many parallel techniques as possible. The addition of paleomagnetic secular variation records and AMS ^{14}C dating can make varve chronology in North America an extremely powerful correlation tool with unparalleled resolution. Based on multiple techniques we present a comprehensive late Pleistocene chronology for events that can be tied to the New England Varve Chronology, especially deglaciation in and around northwestern New England.

GLACIOLACUSTRINE SEDIMENTS AND VARVE CHRONOLOGY

EARLY VARVE INVESTIGATIONS

Silt and clay in the upper Connecticut Valley were first identified as lacustrine sediment in 1818 by Edward Hitchcock (Lougee, 1957). Warren Upham (1878) described the silt and clay as glacial flood deposits and it is not clear whether he recognized them as lacustrine sediment. Seasonal sedimentation patterns in the lacustrine sediment of the Connecticut Valley from Hanover to Woodsville and in the Ammonoosuc Valley were described in detail by Robert Sayles (1919) who may have been the first American to use the term 'varve' in connection with Connecticut Valley sediment. Sayles (1919) compared the Pleistocene varves with

rhythmic units associated with the Squantum "tillite" Member of the Precambrian Roxbury Formation in Boston. Gerard De Geer visited the Champlain Valley as early as 1891 (De Geer, 1921) and later led a Swedish varve expedition in 1920 in which Ernst Antevs, Ebba Hult De Geer, and Ragnar Lidén accompanied him. This expedition introduced Ernst Antevs to the Connecticut Valley where he almost single handedly created the New England Varve Chronology. De Geer's goal was to create a varve chronology in North America that could be compared to varves in Sweden. De Geer (1921) found compelling year-to-year matches and inferred a correlation of varves between New England and Sweden. Varve sections at Woodsville, New Hampshire and Wells River, Vermont were the centerpieces of his interpretations. Antevs (1931, 1935, 1953, 1954) later dismissed De Geer's transatlantic correlations with claims of arbitrary correlations and inconsistencies with ^{14}C ages, an analysis that most modern geologists accept as correct.

THE NEW ENGLAND VARVE CHRONOLOGY — FIRST INSTALLMENT

In 1920 Ernst Antevs began assembling the New England Varve Chronology from varve sections in the Connecticut, Hudson, Ashuelot, and Merrimack Valleys (Figs. 1, 2). He compiled over 4000 consecutive years and published his work in 1922 (Antevs, 1922), an amazing accomplishment for less than two years work. To assemble the chronology Antevs created "normal" curves that are the averaged results from several outcrops and include matching sequences from different drainage basins. This matching or "connection" of individual varve records is necessary to eliminate errors resulting from missing couplets or single varves mistakenly counted as multiple years at one exposure. Except for the tail ends of his entire chronology and two short intervals represented by varves in Massachusetts (NE varves 4324-4341 and 4684-4702), Antevs constructed all his normal curves from varve records at two or more outcrops (Fig. 3). The interval of NE varves 4684-4702 has now been replicated in a core at Amherst, Massachusetts (T. Rittenour, pers. comm.). More than 80% of the normal curves are based on overlapping records from three or more outcrops and as many as 14 outcrops.

Antevs (1922) assembled a continuous lower Connecticut Valley sequence arbitrarily beginning with NE varve 3001 in southern Connecticut and ending just south of Claremont, New Hampshire with NE varve 6277 (Figs. 1, 2). He used varve sections measured by De Geer in the Hudson Valley that matched the Connecticut Valley records to span a gap in the lower Connecticut Valley sequence between NE varves 5600 and 5687 (Fig. 4a, 4b). This interval has now been covered by a measured varve sequence in a 40-m core taken from the Connecticut Valley at Amherst, Massachusetts. From this new core it appears that De Geer over counted his Hudson Valley sequence at one place by 10 years where flood events created a series of couplets that he mistook for annual layers (T. Rittenour and J. Brigham-Grette, pers. comm.). Antevs (1922) was able to match the lower Connecticut Valley sequence with varves in the Ashuelot Valley (NE varves 5687-5733 and 5804-5879, Fig. 4b) and the Merrimack Valley

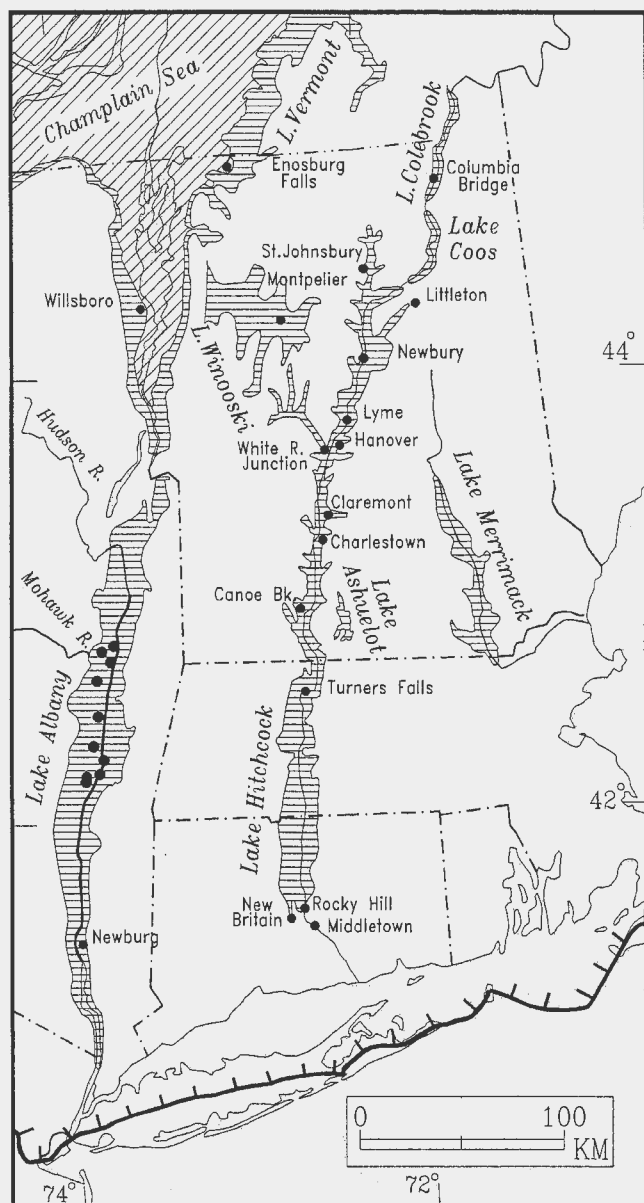


FIGURE 1. Location map of western New England and adjacent New York. The ice front position at the southern end of the map is the maximum extent of Late Wisconsinan ice advance. The Champlain Sea and major glacial lakes relevant to the New England Varve Chronology are shown. The mapped extent of Lake Winoski includes several lake levels in the Winoski Valley (Larsen, 1987b). Unlabeled sites in Lake Albany are locations of varve sections studied by Gerard De Geer (Antevs, 1922).

Carte de localisation de l'ouest de la Nouvelle-Angleterre et de l'état de New York adjacent. L'emplacement du front glaciaire à l'extrémité sud de la carte montre l'extension maximale de l'inlandsis au Wisconsinien supérieur. On montre la Mer de Champlain ainsi que les principaux lacs glaciaires en rapport avec la New England Varve Chronology. Le Lac glaciaire Winoski cartographié (Vermont), comprend différents niveaux lacustres dans la vallée du Winoski (Larsen, 1987b). Les sites non identifiés dans le Lac glaciaire Albany montrent l'emplacement des coupes de varves étudiées par Gerard De Geer (Antevs, 1922).

of eastern New Hampshire (NE varves 5709-5749, 5771-6352, Fig. 4c), which extended the lower Connecticut Valley sequence to NE varve 6352 (Fig. 2).

Beginning in Claremont, and continuing northward into the Passumpsic Valley to St. Johnsbury, Vermont, Antevs (1922) assembled an upper Connecticut Valley sequence (NE varves 6601-7400) that did not appear to overlap his lower Connecticut Valley sequence (Fig. 2). The apparent lack of a match between the upper and lower valley sequences prompted Antevs to create a gap between the two chronologies. The artificial gap (here called the Claremont Gap) was inferred to represent a 600-yr stillstand of the receding ice sheet despite a lack of any field evidence in the Connecticut Valley to suggest such an event. Creation of the Claremont Gap was stimulated by evidence of oscillating ice (till over varves) to the east in the Lake Winnepesaukee region (Antevs, 1922), which needs further investigation. At one locality near Newbury, Vermont Antevs (1922, site 73) counted an additional 1100 couplets (beyond NE varve 7400) that were too thin to measure with a ruler. This section would become the focus of later work in sedimentary paleomagnetism and is the site of new ^{14}C ages and paleomagnetic data discussed later in this paper.

NEW ENGLAND VARVE CHRONOLOGY — SECOND INSTALLMENT

After studying varves in Canada for a few years (Antevs, 1925) Antevs returned to the United States to complete the New England Varve Chronology. Antevs (1928) extended the lower Connecticut Valley sequence back to NE varve 2701 by measuring new sections in Connecticut and matching them with varves in the Hudson Valley at Newbury, New York (Figs. 1, 2). The last two years of his original upper Connecticut Valley sequence were revised and he extended that sequence to NE varve 7750. Antevs also compiled varve records from across northern Vermont and into Québec, which were too short and mostly too young to be matched to his sequence in the Connecticut Valley. However, varves from glacial Lake Winoski, resting directly on till at Montpelier (Fig. 1), matched the upper Connecticut Valley sequence (NE varves 7059-7288; Figs. 2, 4d) and provided a tie to glacial lakes in the Champlain–St. Lawrence drainage system. Other publications by Antevs (1925, 1931) in Canada provide additional chronologies that record deglaciation well into the Holocene.

In this volume in honor of Dick Goldthwait it is worth mentioning that Dick assisted Ernst Antevs in the construction of his varve chronology. Shortly after publishing a paper on varves in the Connecticut Valley (Ridge and Larsen, 1990), the first author received a letter from Dick Goldthwait revealing that his family hosted Ernst Antevs while he completed his work in the upper Connecticut Valley. Dick remembered as a teenager helping Antevs compile varve records on summer evenings in the attic of his family's home in Hanover, New Hampshire.

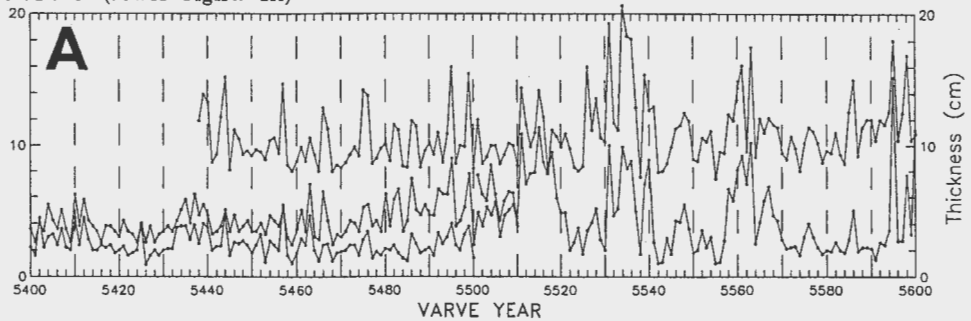
SYSTEMATIC ICE RECESSION IN NEW ENGLAND

Antevs (1922, 1928) was able to use his upper Connecticut Valley sequence to assemble a chronology of deglaciation for the upper valley. At fourteen localities along the axis of the Connecticut and Passumpsic Valleys from Claremont

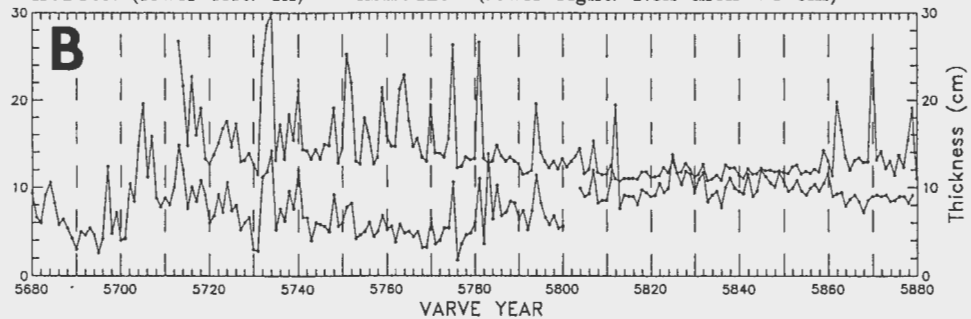
FIGURE 4. The overlap and match of normal curves (Antevs, 1922, 1928) from different drainage basins and glacial lakes. Included on the 200-year plots are normal curves from the Connecticut, Hudson, Ashuelot, and Merrimack Valleys. Many sequences are scaled to fit on the plots and then shifted an increment of the vertical axis (e.g. +6 cm) to allow easier comparison. Varves plotted on the top axis of a graph are excessively thick varves that actually plot well off the graph. Note: The best way to view the varve records is to hold the page at a low angle, looking across the page from bottom to top.

Le chevauchement et la concordance des courbes normales (Antevs, 1922, 1928) dans différents bassins de drainage et lacs glaciaires. Sont comprises dans la reconstitution de 200 ans les courbes normales des vallées du Connecticut, de l'Hudson, de l'Ashuelot et du Merrimack. Plusieurs séquences ont été ajustées puis décalées pour fins de comparaison. Les varves qui touchent la partie supérieure du graphique sont excessivement épaisses et excèdent largement le graphique. Note : la meilleure façon de consulter le relevé est de tenir la page à un angle faible, en la parcourant du bas vers le haut.

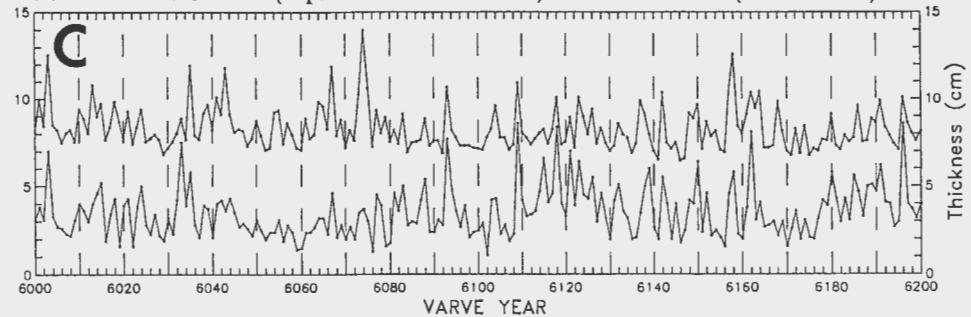
LOWER CONNECTICUT - Mass. (2 lower left records: top 2X then +1 cm, bot. 1X)
 LOWER CONNECTICUT - New Hampshire (upper right: 1/3X then +6 cm)
 HUDSON (lower right: 1X)



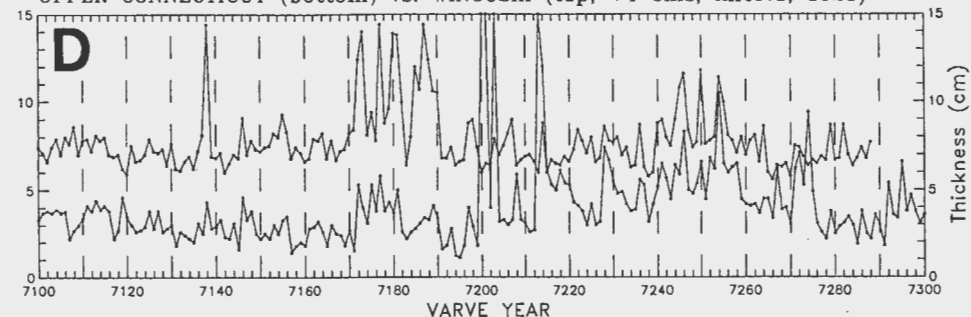
LOWER CONNECTICUT - Vermont (top: 1/2X then +10 cm)
 HUDSON (lower left: 1X) ASHUELÔT (lower right: 1.5X then +4 cm)



LOWER CONNECTICUT (top: 0.6X then +6 cm) vs. MERRIMACK (bottom: 1X)



UPPER CONNECTICUT (bottom) vs. WINOOSKI (top, +4 cms; Antevs, 1928)



of the depth to till or bedrock in the center of a valley it is impossible to unequivocally deny the existence of deeper and older parts of the varve stratigraphy that lie in basins beneath the center of the valley. Antevs (1922) did not report the exact outcrop thickness of basal varves at every site but where he did report them (about half the basal sites) they

reach a thickness of at least 37 cm and are up to 396 cm. These thick varves at basal sections are almost certainly ice-proximal and ice would probably have been within a few kilometers of the site at the time of deposition, thus making the ages of these varves very good approximations of the age of deglaciation.

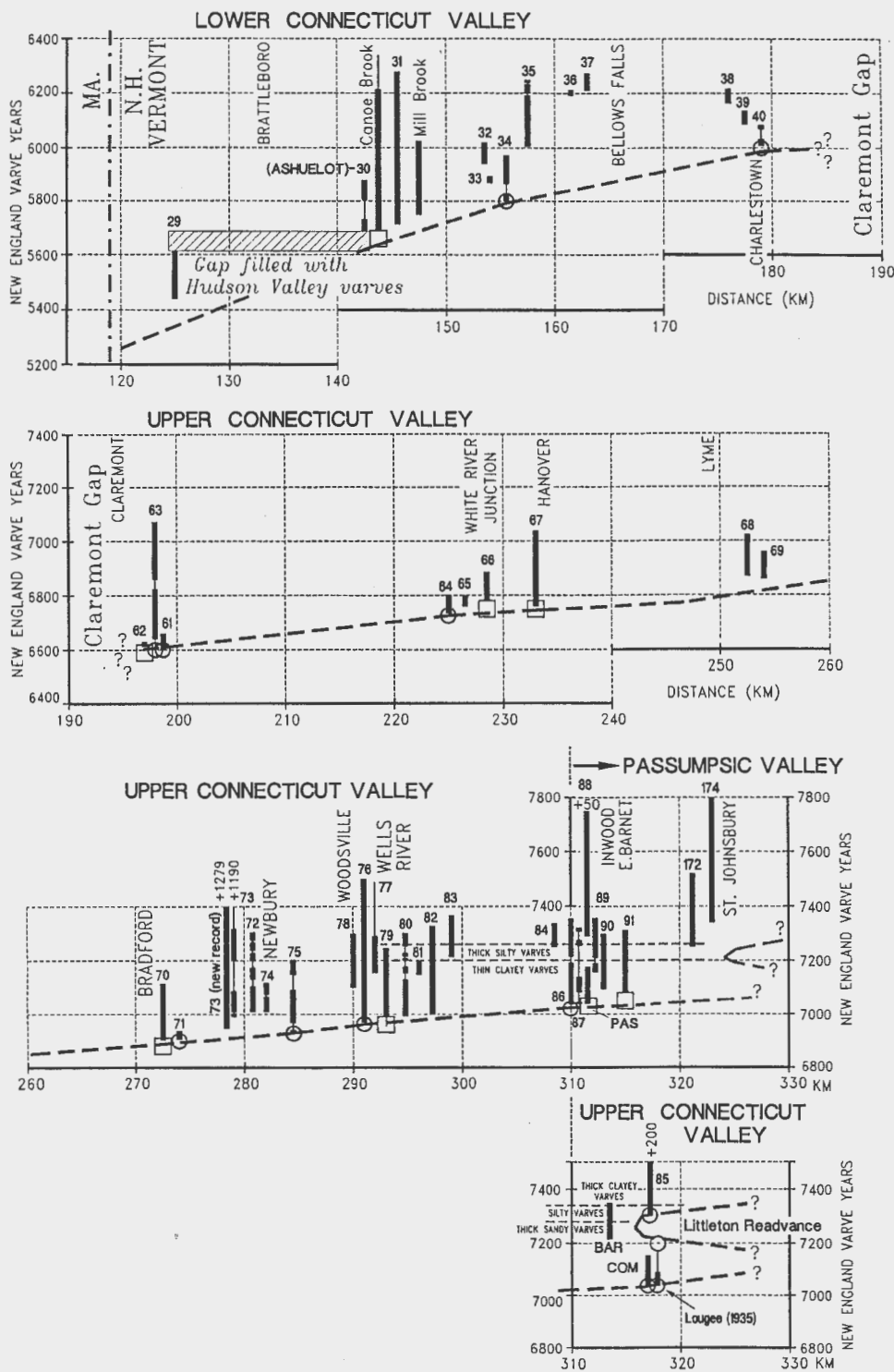


FIGURE 5. Time spans of varve sections in Vermont and New Hampshire along the axes of the Connecticut and Passumpsic Valleys from south (left, top) to north (right, bottom). Distances are from an arbitrary starting point south of Lake Hitchcock in Connecticut. The section of the valley marked Lower Connecticut Valley is the part of the valley covered by the Lower Connecticut Valley varve sequence (Antevs, 1922, 1928). The Upper Connecticut Valley plot from 310-330 km follows the valley eastward from its confluence with the Passumpsic Valley, the latter being the topographic extension of the south to north-trending Connecticut Valley. Thick varve columns are sequences measured and matched to other sections; thin columns were counted only. Numbered sections are from Antevs (1922, 1928) and other named sections are from later works (Lougee, 1935b; Ridge and Larsen, 1990; Ridge *et al.*, 1996; new data). Open circles at the bottoms of sections indicate outcrops where basal varves have been found overlying till, bedrock, or ice-proximal sand and gravel. Open squares indicate section bottoms with thick ice-proximal varves. The dashed line at the bottom of the profiles represents an interpretation of the age of deglaciation (see discussion in text).

Périodes couvertes par les séquences varvaires au Vermont et au New Hampshire le long des axes des vallées du Connecticut et du Passumpsic, du sud (en haut, à gauche) vers le nord (en bas, à droite). Les distances sont calculées à partir d'un point de départ arbitraire au sud du Lac glaciaire Hitchcock, dans le Connecticut. La partie identifiée comme étant la Lower Connecticut Valley est la partie comprise dans la séquence varvaire de la vallée inférieure du Connecticut (Antevs, 1922, 1928). Le relevé de la vallée supérieure du Connecticut de 310 à 330 km suit la vallée vers l'est à partir de la confluence avec la vallée du Passumpsic qui représente l'extension naturelle de la direction sud-nord de la vallée du Connecticut. Les colonnes plus larges de varves représentent des séquences mesurées et assorties à d'autres coupes; les colonnes fines ne

*comprennent que des mesures. Les coupes numérotées sont de Antevs (1922, 1928) et les coupes nommées résultent de travaux ultérieurs (Lougee, 1935b; Ridge et Larsen, 1990; Ridge *et al.*, 1996; données inédites). Les cercles non tramés à la base des colonnes identifient les coupes où les varves couvraient un till, le substratum ou des sables et des graviers proximaux. Les carrés non tramés identifient des coupes où la base comprend des varves épaisses proximaux. Les tirets à la base indiquent l'époque probable de la déglaciation.*

At most places where basal varves have been recorded in sections along the valley side there is very little or no opportunity for the accumulation of thick subsurface varve stratigraphy

that could significantly predate valley side surface exposures. This is especially true in the upper Connecticut and Passumpsic Valleys. At Claremont (Fig. 5), where Antevs (1922) found

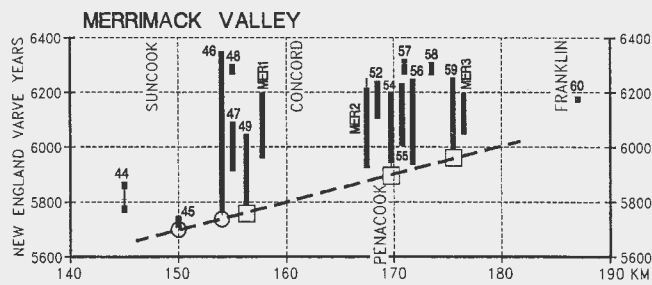


FIGURE 6. Time spans of varve sections along the axis of the Merrimack Valley from south (left) to north (right). Sites marked MER1-3 are new sections. Distance is from an arbitrary point south of the Merrimack Valley in Massachusetts. Symbols and plotting convention are the same as for Figure 5.

Périodes couvertes par les séquences de varves le long de l'axe de la vallée du Merrimack, du sud (à gauche) vers le nord. Les sites identifiés MER1-3 représentent de nouvelles coupes. Les distances sont calculées à partir d'un point arbitraire au sud de la vallée du Merrimack, au Massachusetts. Voir la figure 5 pour les symboles et la méthode utilisée.

his thickest basal varves (up to 396 cm, first 30 average >50 cm), bedrock and till are exposed along the banks of the Connecticut River. The center of the valley appears to be shallow and would have rapidly filled with fine sand, silt, and clay given the documented thickness of varves in surface exposures. At Wells River (Fig. 5) the bridge crossing the Connecticut River to Woodsville sits on bedrock outcrops and basal varve sections occur on till near river level. Basal varves in the Passumpsic Valley (Fig. 5) are exposed down to river level where the river runs across bedrock for 2 km.

In addition to areas of the upper Connecticut Valley where bedrock is shallow basal varves are also exposed above thick sequences of till and preglacial deposits that are now dissected by the Connecticut River. The till and preglacial deposits filled the valley above modern river level at the time of deglaciation and the Connecticut River has dissected the pre-varve deposits in postglacial time making the valley deeper today than during deglaciation. South of Hanover (Fig. 5) the Connecticut River is inset in a bluff that is composed of till and preglacial gravel totaling 22 m capped by basal varves (Larsen, 1987a). On the New Hampshire side of the Comerford Dam in the upper Connecticut Valley (located at about 318-319 km on Fig. 5) over 50 m of till and preglacial gravel sit on bedrock and have been downcut by the modern Connecticut River that today runs on bedrock. Till on both sides of the valley is capped by varves deposited during deglaciation (Ridge *et al.*, 1996; Thompson *et al.*, 1999). During deglaciation the valley was filled with till and preglacial gravel and could not have been a location where a deep varve section accumulated significantly below the elevation of existing exposures along the river. Finally the Littleton-Bethlehem Readvance (discussed later) is recorded in the varve sequence fixing the age of deglaciation near the Comerford Dam and Littleton.

In situations where valley side exposures of basal varves occur next to subsurface basins in the center of the valley it is unlikely that significantly older varve sections reside in

the basins. In situations where deposition along a valley side is delayed sediment is focused to deeper parts of a basin and varves lap up on the valley side. Varves in deep basins in the center of the valley should become thinner as they are traced outward and begin to lap up on till or bedrock on the sides of the valley. Using the bottom varves at Canoe Brook as an example (Figs. 1, 5; NE varve 5685 = 25.5 cm, 5686 = 23.0 cm, 5687 = 30.0 cm, and 5688 = 28.0 cm; Ridge and Larsen, 1990 and new data), valley-side basal varves conservatively have an average thickness of 20 cm. Thicker varves in the center of the valley would fill a 50-m basin in no more than 250 yr. The minimum thickness of basal varves in the center of the valley might better be represented by basal varves in parts of the valley where bedrock is shallow, such as Claremont (average of +50 cm for basal 30 varves, up to 396 cm, Antevs, 1922) or the Passumpsic Valley (45-180 cm; Antevs, 1922). Using these rates, valley filling would occur in 100 yr or less assuming that all valley filling was by varve deposition. However, subaqueous fan deposition at the receding ice front is likely to have filled some portion of the basin with sand and gravel in the first few years after deglaciation. Also, unless a cross-valley profile is very steep complete basin filling would not have to predate valley-side varve deposition and varves can drape the flanks of the basin. Differences in elevation of contemporaneous varves suggest that sedimentation did not closely follow a pattern dominated by the ponding of silt and clay in the center of the valley. All of these characteristics of valley filling during varve deposition suggest that there are not likely to be significant differences between the ages of valley-center and valley-side basal varves in the Connecticut and Passumpsic Valleys of New Hampshire and Vermont. Shallow bedrock in many places and rapid sedimentation rates for ice-proximal varves appear to make basal varves reliable recorders of the age of deglaciation in these valleys.

Critical data from areas with basal varve exposures in the Merrimack Valley do not exist to unequivocally make the same detailed arguments as in the Connecticut Valley. However, there seems to be no reason to expect the Merrimack Valley to behave differently than the upper Connecticut Valley. It has a lower relief than the upper Connecticut Valley, which would allow more draping of varves on valley sides, and basal varves at two sites reach a thickness of 75 cm (Antevs, 1922). In both the Connecticut and Merrimack Valleys basal varves always get younger to the north, a pattern that might not persist if there were differences in age for basal varves located at different cross-valley positions. The conclusion that we reach is that the existing data on basal varve ages in the upper Connecticut and Merrimack basins of New Hampshire and Vermont provide an estimate of the age of deglaciation that does not have significant errors. A similar situation might not exist in the lower Connecticut Valley of Massachusetts and Connecticut where basal varve localities are scarce and the valley is much wider and has larger and deeper subsurface basins. However, a 40-m core down to till taken on the floor of the Connecticut Valley at Amherst has a basal varve age in

exact agreement with ages of basal varves found by Antevs (1922) in nearby valley-side surface exposures (T. Rittenour and J. Brigham-Grette, pers. comm.).

The onlapping varve sequences (Figs. 5, 6) record systematic recession of ice from south to north as did Antevs' original analysis that was in the form of a map showing the varve age of deglaciation across New England (Plate VI, Antevs, 1922, compiled with J.W. Goldthwait). However, within a decade of Antevs' original work, Richard Foster Flint (1929, 1930, 1932, 1933) denied systematic ice recession in favor of regional ice stagnation, and more importantly, he also created doubts as to the validity of the varve chronology (Flint, 1930). Flint's later publications (1932, 1933) were apologetic and supportive of the varve chronology as a dating tool, but varve chronology was not viewed with the credibility it had before Flint's criticism. In the 1950's some of the first ^{14}C ages obtained in New England were erroneously interpreted by Flint (1956) to constrain the age of Lake Hitchcock to about 2500 yr. Despite Antevs' (1962) objections these ^{14}C interpretations caused further erosion of confidence in the New England Varve Chronology. Although referenced in the first of Flint's (1947) textbooks on glacial geology, his later two textbooks (Flint, 1957, 1971), used by most students of glacial geology in North America for two decades, have no reference to Antevs' (1922, 1928) work on varves in New England. More recent studies of varves in New England have focused on the sedimentology of the varve sequence, in particular depositional processes and facies changes associated with the varves, and have not addressed issues of chronology (Ashley, 1972, 1975; Ashley *et al.*, 1982). While most glacial geologists working in New England over the last half century have ignored varve chronology as a viable chronologic tool, many investigators (Lougée, 1935b; McNish and Johnson, 1938; Johnson *et al.*, 1948; Verosub, 1979a, 1979b; Thomas, 1984; Ridge and Larsen, 1990, Ridge *et al.*, 1995, 1996; Levy, 1998; T. Rittenour, pers. comm.; A. Werner, pers. comm.) have been able to match new varve sections to Antevs' (1922, 1928) chronology.

OVERLAP OF THE LOWER AND UPPER CONNECTICUT VALLEY VARVE SEQUENCES

Attempts to match Antevs' (1922) lower and upper Connecticut Valley sequences, using visual inspection and mathematical correlation techniques with the aid of a computer, have revealed one potential correlation of the two sequences. With this overlap in place (lower NE varve 6012 = upper NE varve 6601) the two varve records have a greater visual resemblance in terms of relative peak heights and peak position than at any other overlap (Figs. 2, 7). This correlation is not represented by a match that is as compelling as any proposed by Antevs (1922, 1928) when he compiled his sequences in the Connecticut Valley. It is also less compelling than correlations between varves from different glacial lakes in New England (Fig. 4). New ^{14}C ages and paleomagnetic data are used later in this paper to support the correlation of the lower and upper Connecticut Valley sequences. The proposed correlation of the lower and upper valley sequences eliminates the necessity for a stillstand of the receding ice front near Claremont and creates a continuous rapid recession of ice from southern to northern Vermont and New Hampshire.

Three obstacles appear to have prevented a clear correlation of the lower and upper Connecticut Valley sequences. First, the two sequences have varves with greatly different couplet thickness. The first 30 couplets of the upper sequence average more than 50 cm and couplets in the second 100 yr still have an average of about 20 cm (Fig. 7). The corresponding part of the lower Connecticut Valley sequence has couplets with an average thickness of about 3 cm. Second, the thick varves in the beginning of the upper sequence are more likely to have a thickness that reflects local ice-proximal sedimentation (Ridge and Larsen, 1990), rather than regional weather patterns, making them less useful for regional correlation. The varve sections used to compile the first 275 varves of the upper Connecticut Valley sequence (Fig. 5) are all located close to the mouths of rivers near Claremont (Sugar River) and White River Junction, (White and Mas-

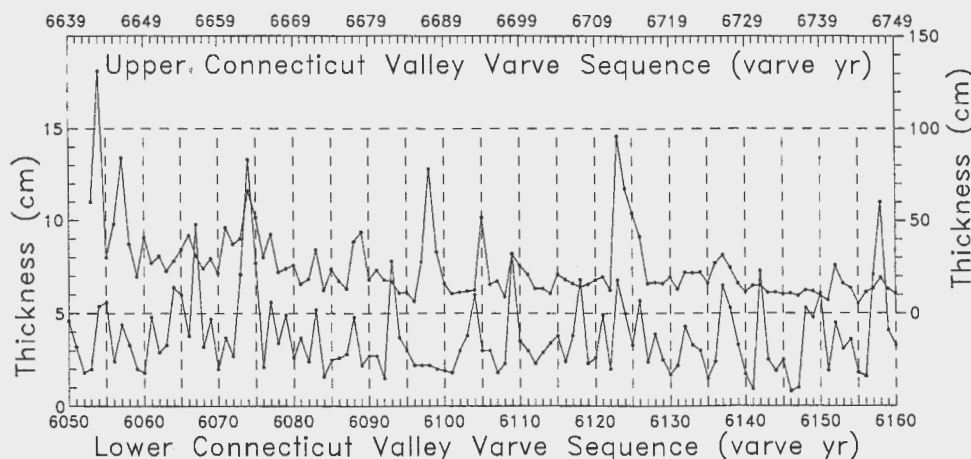


FIGURE 7. Correlation of the lower (bottom) and upper (top) Connecticut Valley normal curves of Antevs (1922). Lower NE varves 6053-6160 are correlated to upper NE varves 6642-6749. Note the different scales used to plot the sequences.

Corrélations proposées entre les courbes normales des vallées inférieure et supérieure du Connecticut de Antevs (1922). Les varves NE (New England) 6053-6160 (vallée inférieure) correspondent aux varves 6642-6749 de la vallée supérieure. Noter les différentes échelles appliquées.

coma Rivers). Varves deposited at these positions were influenced by large point source supplies of sediment and likely record floods due to localized precipitation events or the release of water from ice-dammed lakes in tributaries. Varves near Claremont have the added complication of possibly being deposited by northward flowing bottom currents coming from the Sugar River and resulting in localized ponding of sediment. Finally, varves used to construct the last 50 years of the lower Connecticut Valley sequence in the Connecticut Valley and the last 50 years of the Merrimack Valley sequence (both overlap the upper Connecticut Valley varves) are varves influenced by deltaic sedimentation. Varve thickness in these sequences is probably dominated by local sedimentation processes and not regional weather patterns. Although couplets in both the upper and lower Connecticut Valley sequences are annual layers, couplet thickness that does not dominantly reflect a regional weather pattern will negate its easy use for regional correlation.

LAKE STAGES IN THE UPPER CONNECTICUT VALLEY

GLACIAL LAKE HITCHCOCK

The first compilation of glacial lake stages in the Connecticut Valley was by Richard Lougee (1935a), who named Lake Hitchcock after Prof. Edward Hitchcock of Amherst College. Lougee's Lake Hitchcock (1935a, 1939, 1957) extended from Middletown, Connecticut (Fig. 1) northward into New Hampshire. Lougee maintained that the spillway for the lake was near Middletown, despite recognition of another more viable spillway at New Britain (Loughlin, 1905; Flint, 1933; Jahns and Willard, 1942; Fig. 1). A drift dam for the lake was also identified at Rocky Hill, Connecticut (Flint, 1933, 1953), thus completely eliminating the need to extend the lake further south. Modern studies have confirmed and refined the early history of lake levels associated with the development of the New Britain spillway and the failure of the Rocky Hill dam (Hartshorn and Colton, 1967; Koteff *et al.*, 1987; Koteff and Larsen, 1989; Stone *et al.*, 1991; Stone and Ashley, 1992; Stone, 1995). Lougee recognized deltas from Lake Hitchcock as far north as Lyme, New Hampshire (Fig. 1), which he inferred to represent the northward extent of the lake when its dam was breached (Lougee, 1939, 1957).

LOUGEE'S LAKE UPHAM

Lougee's (1939, 1957) interpretation of terraces in the Connecticut Valley near Lyme and Hanover (Fig. 1), as remnants of lake floor from Lake Hitchcock, led him to conclude that the breaching of the Lake Hitchcock dam allowed water to fall 30 m giving way to Lake Upham. Lake Upham was named after Warren Upham who was an early investigator of Connecticut Valley terraces and eskers. Lougee (1957) inferred that Lake Upham drained by way of a channel that was cut across lake floor deposits south of Charlestown, New Hampshire and was graded to a bedrock ledge at Turners Falls, Massachusetts. Lougee extended Lake Upham

northward to St. Johnsbury where he proposed a hinge line and an unrealistically steep water plane (>1.6 m/km) to intersect deltas at high elevations in the upper Connecticut and Ammonoosuc Valleys (Fig. 8). Lougee (1957) referred to precise survey data of strandline features collected in the 1920's to support his proposed lake levels, but these data were not published and have not been found in the archives at Dartmouth College (W. Thompson, pers. comm.).

MODERN STUDIES OF ISOSTASY AND LAKE HITCHCOCK DRAINAGE

Leveling surveys by Jahns and Willard (1942) in Massachusetts, that have been greatly refined and expanded to northern New Hampshire and Vermont (Koteff and Larsen, 1989), indicate that the level of Lake Hitchcock was controlled by the New Britain channel in Connecticut. Koteff and Larsen also identified deltas defining a flat, tilted (0.9 m/km, up at 339°) water plane for Lake Hitchcock that extends northward to at least Woodsville and Littleton (Fig. 1), thus refuting Lougee's (1939, 1957) formation of Lake Upham while the receding ice margin was at Lyme. Controversy still exists regarding when Lake Hitchcock abandoned the New Britain spillway and drained at a lower level. ^{14}C ages of plant debris from Connecticut in deposits that represent the drainage of Lake Hitchcock have been interpreted to indicate that drainage occurred at about 13.5 ^{14}C ka (Stone *et al.*, 1991; Stone and Ashley, 1992; Stone, 1995). It has been suggested that perhaps Lake Hitchcock completely drained at about 13.5 ^{14}C ka while the receding ice front was in central Massachusetts at the Holyoke Range (Ashley, 1995). The Holyoke Range is then thought to have served as a barrier allowing the continued impoundment of water to the north during subsequent ice recession. However, a thick package of varves immediately south of the Holyoke Range was deposited almost up to the water plane level of Lake Hitchcock (Werner, 1995). These varves indicate that Lake Hitchcock persisted long after recession of ice north of the Holyoke Range, which does not seem to have served as a dam or spillway.

Drainage of Lake Hitchcock while the receding ice front was in northern New Hampshire and Vermont has been inferred by the identification of Lake Hitchcock deltas as far north as Littleton and Woodsville (Larsen and Koteff, 1988; Koteff and Larsen, 1989). In central Vermont a late drainage for Lake Hitchcock is indicated by Lake Hitchcock deltas found from White River Junction (Fig. 1) northwest into the White River Valley that became a long embayment of Lake Hitchcock (Larsen, 1987a). Lake Hitchcock deltas in the upper White River Valley were dissected by spillway drainage from Lake Winooski (Larsen, 1987b) indicating that Lake Hitchcock drained while Lake Winooski was in existence (Larsen, 1984, 1987a). Antevs' (1928) varve record from Lake Winooski (Fig. 4d) matches the upper Connecticut Valley varve sequence (Antevs, 1922) and indicates that Lake Hitchcock may have drained sometime during the deposition of NE varves 7059-7288 (Fig. 2). By this time ice had receded to the vicinity of St. Johnsbury and Littleton (Fig. 5).

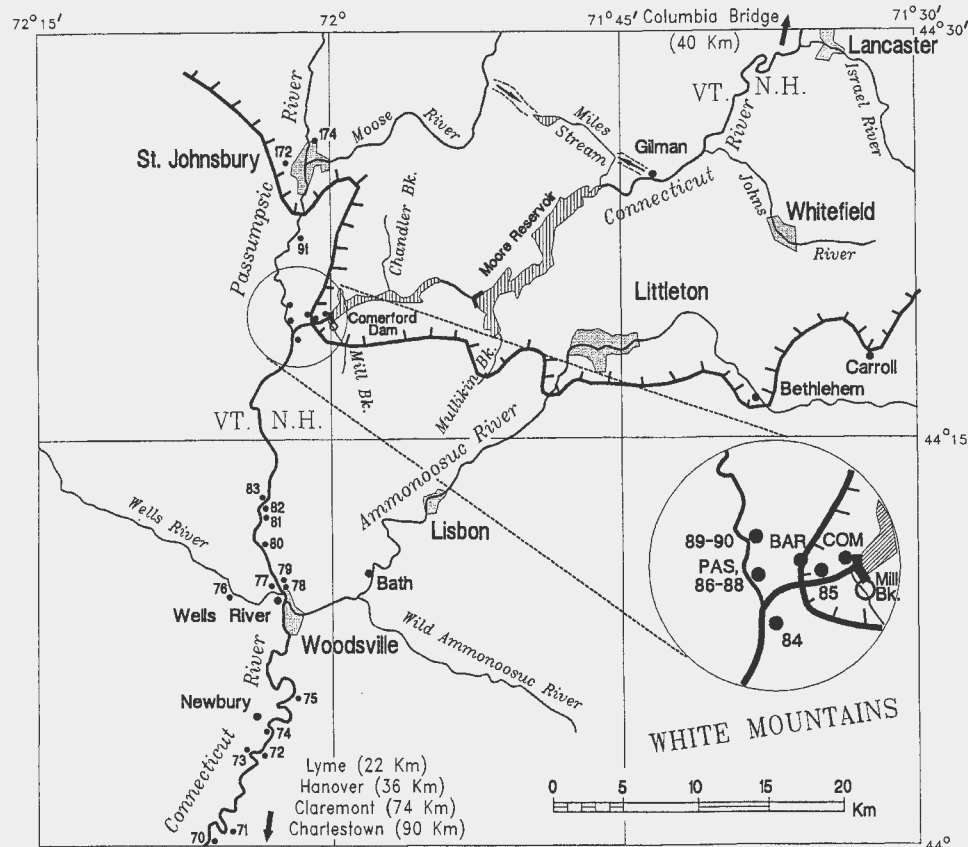


FIGURE 8. Location map of the upper Connecticut Valley near its confluence with the Passumpsic Valley northwest of the White Mountains. The numbered sites are varve sections studied by Antevs (1922, 1928). New sites are listed as BAR, COM, and PAS. The ice front position is the approximate limit of the Littleton-Bethlehem Readvance.

Carte de localisation de la vallée supérieure du Connecticut près de la confluence avec la vallée du Passumpsic, au nord-ouest des White Mountains. Les sites de varves numérotés ont été étudiés par Antevs (1922, 1928). Les nouveaux sites sont identifiés par les lettres BAR, COM et PAS. Le front glaciaire donne la limite approximative de la récession de Littleton-Bethlehem.

The conflict between the drainage of Lake Hitchcock in Connecticut at 13.5 ^{14}C ka (Stone and Ashley, 1992; Stone, 1995) and younger ages for Lake Hitchcock in the upper Connecticut Valley (discussed later in this paper) has not yet been resolved. It is here suggested that perhaps the macrofossils dated at 13.5 ^{14}C ka used to infer the age of beds representing the drainage of Lake Hitchcock in Connecticut are from reworked older organic sediment. At the very least the upper Connecticut Valley appears to have had a lake in it during deglaciation that exactly coincides with the northward projection of the Lake Hitchcock water plane from Connecticut and Massachusetts (Koteff and Larsen, 1989). The exact position of the receding ice sheet when this lake in the upper valley drained has not yet been determined and has been difficult to study. Deltas from Lake Hitchcock are seldom preserved and they are mostly confined to tributary valleys as a result of rapid ice recession, the short life span of Lake Hitchcock in the upper Connecticut Valley, and the dissection and trimming of these features by later drainage.

POST-HITCHCOCK LAKES

Throughout the Connecticut Valley there is pervasive evidence for lake levels below the level of Lake Hitchcock. Drainage of Lake Hitchcock in Connecticut allowed a river to form high terraces on an exposed lake bed (Stone and Ashley, 1992), and when combined with isostatic depression at the time, would have allowed at least low level lakes to persist further north in the valley. The accordance of deltas in

the valley on water planes approximately parallel to and about 8-10 m lower than the projected Lake Hitchcock water plane (Larsen and Koteff, 1987a; Koteff and Larsen, 1989) suggests that a discrete lower lake level formed prior to significant isostatic tilting in the area. This lake is well represented by meteoric deltas in southern New Hampshire and Vermont at the mouths of the Cold and Saxtons Rivers (Ridge, 1988). This lake level also appears to be represented by ice-contact deltas in the Chandler Brook and Ammonoosuc Valleys of the upper Connecticut Valley (Ridge *et al.*, 1996; Fig. 8). It is clear that Lougee (1939, 1957) did not recognize these deltas as representing a separate lake level. In some places Lougee tried to associate the deltas with Lake Hitchcock or created a hinge line and steep upper end of the Lake Upham water plane that intersects these features.

In the upper Connecticut Valley additional deltas record water levels that are 20 m or more below the level of Lake Hitchcock (Koteff and Larsen, 1989). Lake levels in this range are recorded by deltas and terraces near Hanover and Lyme (Fig. 1) that were used by Lougee, 1939, 1957 to define his Lake Upham. Evidence for low lake levels has been found further north (Koteff and Larsen, 1989) where they appear to be represented by terraces at Chandler Brook (Ridge *et al.*, 1996; Fig. 8). Perhaps these features represent Lougee's Lake Upham without its hinge line and steep profile to the north. Regardless, ice-contact deltas in the Chandler and Ammonoosuc Valleys at an elevation about 8 m below

the projected Lake Hitchcock water plane indicate that lower lakes could not form until ice had receded north of St. Johnsbury and Littleton for the final time.

While postglacial lake levels in the upper Connecticut Valley have been difficult to document the persistence of lakes in the valley for at least 1600 yr after deglaciation is easily documented. Antevs counted about 1500 varves at Newbury that begin about 100 yr after deglaciation (Antevs, 1922, site 73; Fig. 8). A reexamination of this section (discussed later) reveals another 179 varve years that were not counted in the 1920's. There also appears to be a thick package of thin varves in the Ammonoosuc Valley that represents a long-lived lake (Sayles, 1919; Billings, 1935).

GLACIAL LAKES COOS AND COLEBROOK

North of Littleton in the Connecticut Valley Lougee (1939) found evidence for glacial lakes extending to Québec at higher levels than the modern projected water planes for Lake Hitchcock (Koteff and Larsen, 1989). The southern of these lakes, Lake Coos, and further north Lake Colebrook require blockage of the Connecticut Valley and diversion of spillway drainage. Lougee (undated) inferred that thick till at Fifteen Mile Falls (Comerford Dam, Fig. 8) provided a dam and stable spillway for lake Coos. However, till at this location is overlain by non-resistant silt and clay well below the projected level of even Lake Hitchcock thus eliminating it as a potential dam or spillway for Lake Coos. A better alternative is a channel at Gilman, Vermont (Ridge *et al.*, 1996) where water could have been diverted to Miles Stream before returning to the Connecticut Valley (Fig. 8). The Connecticut Valley at Gilman has a narrow constriction where till probably blocked its modern drainage path and diverted water into the Gilman channel. Lake Coos deltas identified by Lougee (undated) at the mouths of the Johns and Israel Rivers (Fig. 8) appear to be graded to a lake level compatible with the Gilman spillway. Lake Colebrook was a water body separate from Lake Coos because the floor of the Connecticut Valley and varves from Lake Colebrook at Columbia Bridge (Miller and Thompson, 1979) are above the projected water plane for Lake Coos, even with a steep isostatic tilt of 1.0 m/km.

VARVES AT NEWBURY, VERMONT: FLOOD EVENTS AND LAKE LEVEL CHANGE

In 1997 the first author and his students at Tufts University were able to locate a varve section about 50 m downstream from Antevs' original exposure of varves at Newbury (1922, site 73; Fig. 8) where Antevs counted 1500 couplets beginning at NE varve 6990. Antevs measured couplets up to NE varve 7316 but only counted the remaining couplets because they were too thin and indistinct to measure accurately in the field. We have collected samples of the entire section in two sets of overlapping PVC cores (7.6 cm id, 60 cm long) for measurement and analysis of the varve sequence. Our count was done on cores that were partially dried to improve the color contrast between clay and silt beds. Varve measurement was done with the aid of magnified video images and computer image software that allowed us to make measurements and assemble data files from the images. The new analysis of

the section starts at NE varve 6944, 46 yr below Antevs' measured section. We have been able to match couplets at the exposure with NE varves 6944-7510 and have counted upward from that point to NE varve 8679 (+35/-20) after accounting for uncertainties in the interpretation of annual couplets (179 yr beyond Antevs' count).

In addition to providing an extension of the New England Varve chronology and new ^{14}C ages (discussed later) the Newbury section has a record of abrupt changes in varve thickness and lithology that appear to represent flood events and drops in lake level. Flood events are represented by the abrupt appearance of a few excessively thick couplets, but varve thickness returns to pre-flood thickness within a few years. Drops in lake level are also marked by the abrupt appearance of thick couplets, although usually not as extreme, but lake level drops create thickness and lithologic changes in the varve sequence that persist for many decades or centuries after the event. Persistent changes are caused by increased erosion as stream valleys and the glacial meltwater system adjust to falling base level and deltas and lake floor deposits are exposed to erosion. Also, increased sediment volumes delivered to the lake are distributed across a lake floor surface area that has decreased in size.

Beginning with NE varve 7200 at Newbury there is a 14-yr interval of varves with exceedingly thick couplets that can be found over a distance of 45 km (NE varves 7200, 7202, 7203, and 7213 on Figs. 9a, 9b). Varves in this interval separate thin varves (<1 cm) with silty summer partings below from thicker (2-3 cm) varves that have distinct silt and fine sand summer beds. The same pattern occurs in contemporaneous varves at Wells River near the mouth of the Ammonoosuc Valley where the extremely thick varves reach a thickness of 76 cm (Antevs, 1922). At Wells River the 14-yr varve package separates varves with an average thickness of 2 cm below from varves averaging 5 cm above the interval. This event was also found by Antevs (1922) further north in the Passumpsic Valley near St. Johnsbury (Fig. 8). The exceedingly thick varves occur at about the time that lakes in the upper Ammonoosuc Valley east of Littleton (Lougee, 1940; Thompson *et al.*, 1996, 1999) catastrophically drained to the Connecticut Valley in response to the recession of ice from the Bethlehem Moraines. The Ammonoosuc Valley floods may represent the largest release of lake water impounded in a tributary during the recession of ice in the upper Connecticut Valley. The lithologic and thickness changes that persist after this event (after NE varve 7213, Fig. 9b) also indicate that a drop in lake level occurred in the Connecticut Valley. It is suggested here that, if Lake Hitchcock drained while receding ice was in the northern Connecticut Valley, this event may represent the initial drop in the level of Lake Hitchcock. More speculatively, the Ammonoosuc Valley floods may have provided a triggering mechanism that facilitated the initial failure of the Rocky Hill dam in Connecticut. Alternatively, the lowering of Lake Hitchcock in the lower Ammonoosuc Valley may have facilitated the catastrophic release of lake water in the upper Ammonoosuc Valley by suddenly increasing hydraulic gradients across an ice dam. A resolution of the problem of when Lake Hitchcock drained will be needed to confirm or deny these possibilities.

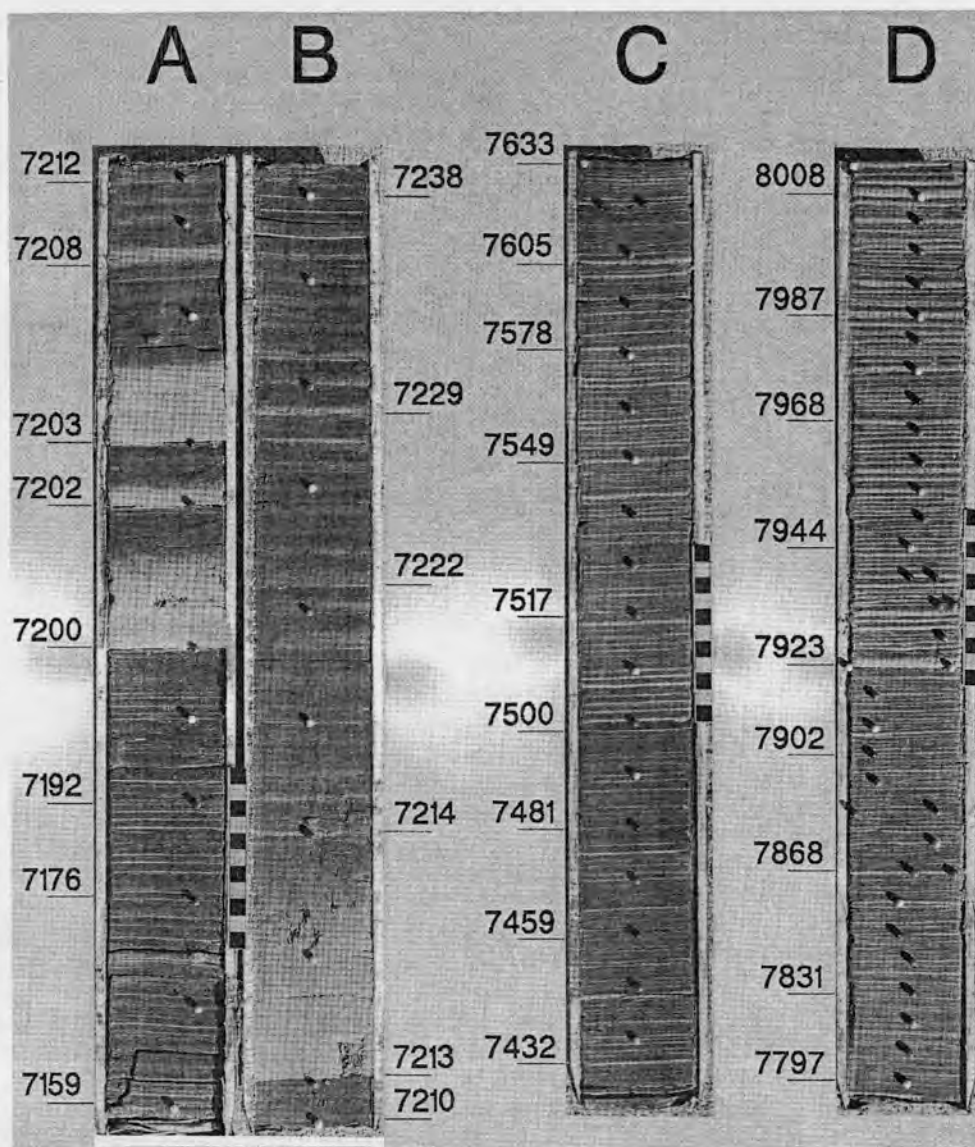


FIGURE 9. Flood and drainage events in varve cores (7.6-cm diameter) from Newbury, Vermont. Scales show centimeters. Pins in the core mark the boundaries of video images. All lines indicating NE varve numbers are positioned at the bottoms of numbered couplets that are either prominent varves or varves marked by pins. A and B) Varves 7159-7238 show a series of flood events in NE varves 7200, 7202, 7203, and 7213. Evidence for a simultaneous drop in lake level comes from the permanent 3 to 5-fold increase in couplet thickness that persists after the flood events and commences at NE varve 7214. C) NE varves 7428-7633 in which NE varve 7500 marks a drop in lake level. Note the sudden increase in thickness at NE varve 7500 and the more conspicuous summer silt beds that persist for decades after this event. D) NE varves 7797-8010 in which sudden and permanent increases in thickness and silt content at NE varve 7923 appear to represent a drop in lake level and possibly a flood event.

Les marques d'inondation et d'écoulement dans les carottes de varves (7,6 cm de diamètre) de Newbury, au Vermont (échelle en cm). Les repères dans les carottes donnent les limites des images vidéo. Toutes les lignes se rapportant aux numéros de varves sont situées à la base des doubles feuillets qui sont soit en saillie, soit identifiés par des repères. A et B) La séquence de 7159 à 7238 montre une série d'inondations dans les varves 7200, 7202, 7203 et 7213. La baisse simultanée des niveaux lacustres est signalée par l'augmentation de 3 à 5 fois l'épaisseur des feuillets qui persiste

après les inondations et qui commence avec la varve 7214. C) Dans la séquence de 7228 à 7633, la varve 7500 montre une baisse du niveau lacustre. Noter la soudaine augmentation dans l'épaisseur à la varve 7500 et les feuillets d'été plus visibles même plusieurs décennies après l'événement. D) Dans la séquence 7797 à 8010, on observe une augmentation soudaine et permanente dans l'épaisseur et la teneur en silt à la varve 7923 qui semble traduire une baisse du niveau lacustre et peut-être une inondation.

Beginning at NE varve 7500 is another abrupt change in varve thickness and lithology that appears to represent another drop in lake level. At Newbury this event is marked by a sudden increase in average varve thickness from about 0.2-0.3 cm to 0.6 cm that persists for about 10-15 years (Fig. 9c). Varves below this interval have very thin silty summer partings and very clayey winter beds. Varves above eventually become slightly thinner, but have slightly thicker and sandier summer beds that are more conspicuous in a partially dried core than below NE varve 7500 (Fig. 9c). In the Passumpsic Valley 45 km to the north (Fig. 8) this event also occurs in the varve stratigraphy with NE varves 7500-7515 being about 5-10 cm thick and varves below less than 1 cm thick (Antevs, 1928). It is suggested that this event

may represent the drop of water levels in the Connecticut Valley from 8-10 m below the water plane of Lake Hitchcock to about 20 m below. A similar more pronounced event occurs later in the Newbury section at NE varve 7923 (Fig. 9d) and may represent another drop in lake level, perhaps down to 30-40 m below the Lake Hitchcock water plane. It is not possible to unequivocally relate these events to specific lake levels and one must also consider flood events caused by the failure of dams for Lakes Coos and Colebrook. However, the events at Newbury do represent important marker horizons in the varve stratigraphy. If they can be related to specific lake levels or flood events they will provide an exact chronology for drainage events in the upper Connecticut Valley.

THE LITTLETON-BETHLEHEM READVANCE

THE BETHLEHEM MORAINES AND VARVE STRATIGRAPHY

Prior to 1910 morainic topography from Bethlehem to Littleton, New Hampshire was interpreted to be the result of northward flowing valley glaciers from the White Mountains at the end of the last glaciation (Agassiz, 1870; Hitchcock, 1878; Upham, 1904; see Thompson *et al.*, 1996, 1999; Thompson, 1999). James W. Goldthwait (1916) re-interpreted the Bethlehem Moraines as ice-marginal deposits built at the southern margin of a receding continental ice sheet. In the Connecticut Valley at the Comerford Dam site (Figs. 5, 8) Antevs (1922) found a basal varve resting on till (NE varve 7305, site 85) that was about 300 yr younger than basal varves resting on bedrock only 3 km to the west in the Passumpsic Valley (NE varve 7010, site 86). Antevs (1922) inferred that the apparent delay in deglaciation in the Connecticut Valley represented a stillstand of ice that was the westward equivalent of the Bethlehem Moraines. He also associated the stillstand with a moraine at St. Johnsbury in the Passumpsic Valley.

COMERFORD DAM CONSTRUCTION SITE

During construction and subsurface investigation for the Comerford Dam, Irving B. Crosby (1934a, 1934b) found pervasive two-till stratigraphy in borings and a large bluff exposure near the foot of the proposed Comerford Dam along Mill Brook (also called Smith Brook) in New Hampshire (Fig. 8). Crosby interpreted the upper of his two tills as representing a readvance. With additional fieldwork in the area he extended the Bethlehem Moraines to 7 km west of Littleton to include ice-marginal deposits in the Mullikin Brook valley. Careful reexamination of the sections on Mill Brook have revealed a more complete stratigraphy than seen by Crosby (Ridge *et al.*, 1996; Thompson *et al.*, 1999). On a higher bank about 250 m up stream from Crosby's section is an exposure of three till units. The basal till unit in this new exposure is also the basal unit seen by Crosby and except for a lack of any weathering it has all of the characteristics of a pre-late Wisconsinan 'lower' till unit seen across much of New England (Koteff and Pessl, 1985; Newman *et al.*, 1990; Oldale and Colman, 1992). The middle till unit in the new exposure appears to be the equivalent of Crosby's upper till which he found overlain by clay (Crosby, 1934b). Crosby did not see the upper till of the new exposure because the top of his section was eroded during the development of a stream terrace that caps his section (Ridge *et al.*, 1996; Thompson *et al.*, 1999). All till units at Mill Brook have the potential of being Late Wisconsinan but this seems especially true of the upper two tills. Stratified deposits separate the upper two till units and the tills represent separate ice sheet oscillations.

Varve sections stratigraphically above Crosby's (1934a, 1934b) upper till unit were exposed on both sides of the Connecticut Valley during the Comerford Dam construction and measured by J.W. Goldthwait and Dick Lougee. All of the sections were matched to the New England Varve Chronol-

ogy (Antevs, 1922, 1928) and were reported by Lougee (1935b) to match varves at Antevs' site 85 (NE varves 7300-7400; Figs. 5, 8), thus postdating any proposed readvance in the area. Lougee did not publish the results or locations of these varve measurements. However, Lougee (1935b) did publish the results of varve measurements from a section in Vermont at the Comerford Dam (300 m east of site 85, Fig. 5) in which 119 varves were sandwiched between a lower stony till and an upper clayey "material resembling till". Despite minor deformation of the varves, which Lougee attributed to overriding ice, he measured and matched the bottom 52 varves with NE varves 7036-7087 of the Passumpsic Valley (Figs. 10a, 10b). This varve section appears to record the time between the initial recession of ice (NE varve yr 7036) and ice readvance (referred to by Thompson *et al.*, 1999 as the Littleton-Bethlehem Readvance) that arrived no earlier than NE varve yr 7154. Based on Antevs' (1922) earlier results final ice recession occurred no later than NE varve yr 7305 (site 85, Fig. 8).

J.W. GOLDTHWAIT'S GREAT RETRACTION

For reasons that today seem inexplicable, given his original vivid descriptions of the topography of the Bethlehem Moraines and recent field observations (Thompson *et al.*, 1996, 1999; Thompson, 1999), J.W. Goldthwait (1938) recanted his interpretations of the moraines as prominent ice-marginal features produced by an active ice sheet (Goldthwait, 1916). Seemingly under the influence of Flint (1929, 1930, 1932, 1933), and to the displeasure of Lougee (1940), Goldthwait reinterpreted the moraines as stratified deposits created by regional stagnation of the last ice sheet. Goldthwait's retraction, along with the support of Flint, and doubts created in the 1930's regarding the validity of the New England Varve Chronology, diminished the significance of the Bethlehem Moraines as ice-front positions.

NEW OBSERVATIONS NEAR THE COMERFORD DAM

We have been able to relocate Lougee's (1935b) Comerford Dam section, or more likely a very similar nearby outcrop (COM on Figs. 5, 8), and again matched the varves to Antevs' (1922) chronology (Ridge *et al.*, 1996; new data). The varve sequence begins with NE varve 7036 overlying till and continues to NE varve 7154 at the top of the section (Fig. 10b). Although our section ends at precisely the same couplet as the top of Lougee's (1935b) section we were unable to find the clayey till-like material visible on Lougee's (1935b) photograph of his exposure. Our exposure was truncated in the final stages of dam construction and NE varve 7154 is overlain by 0.5 m of sandy artificial fill. Varves at our section exhibit wavy bedding and are more compact than the varves in other sections in the area. However, it is not clear whether the deformation is due to overriding ice or mass movement and whether the compaction represents loading by ice.

Further support for the Littleton-Bethlehem Readvance was found in new varve sections about 1.5 km west of Lougee's Comerford Dam section along a small ravine draining south to the Connecticut River (Ridge *et al.*, 1996). Varves at the new site, which is here called the Barnet section (BAR on

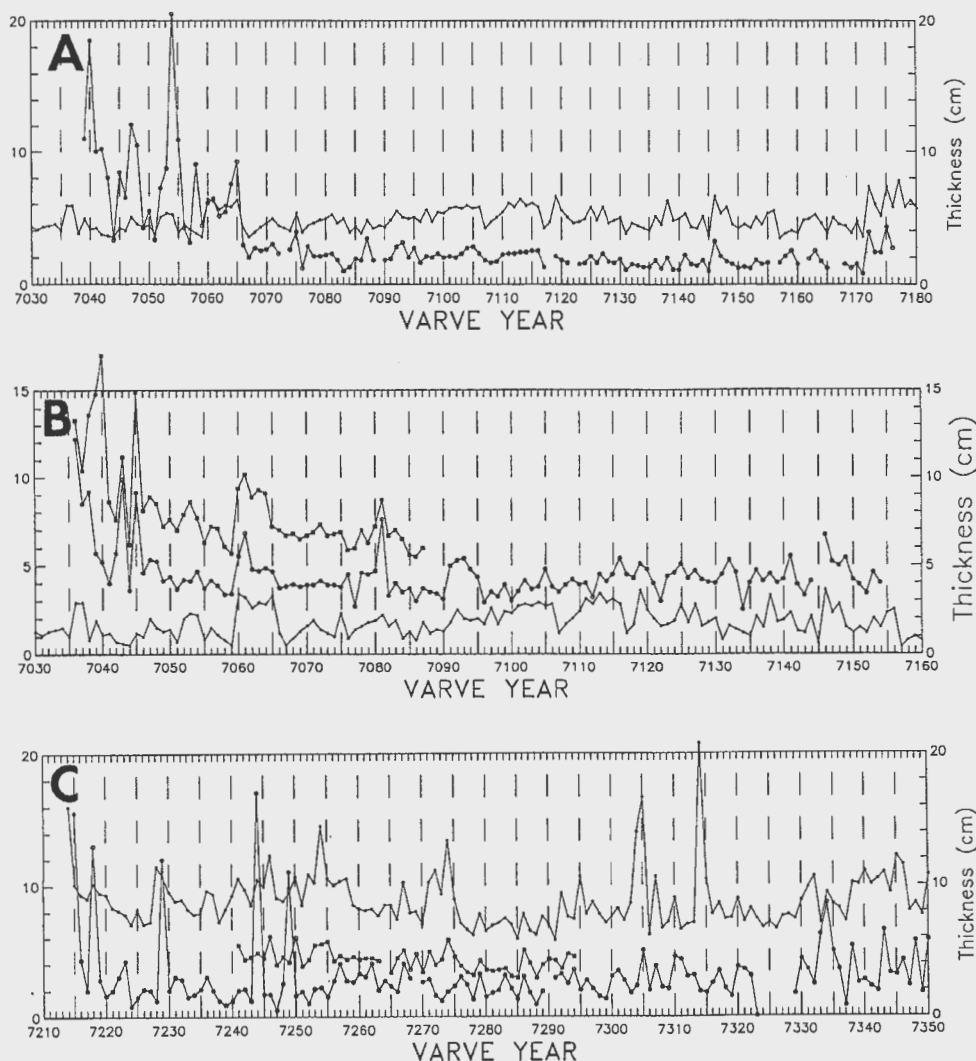


FIGURE 10. Correlation of varve measurements in the upper Connecticut and Passumpsic valleys (Figs. 5 and 8) with the New England Varve Chronology (Antevs, 1922). All plots are at the same scale but some have been shifted as indicated. A) Section PAS (bottom) vs. NE varves 7030-7180 (top, +2 cm). B) Comerford Dam sections in Vermont of Lougee (1935b; at COM on Fig. 8; top, +4 cm) and new composite section (COM on Fig. 8; middle, +2 cm) vs. NE varves 7030-7160 from the Passumpsic Valley (bottom, -1 cm). C) Barnett sections, BAR (bottom; middle, +2 cm) vs. NE varves 7210-7350 (top, +4 cm).

Correspondance entre les mesures des varves des vallées supérieures du Connecticut et du Passumpsic (fig. 5 et 8) et celles de la New England Varve Chronology (Antevs, 1922). Toutes les données sont à la même échelle, mais certaines ont été décalées tel qu'indiqué. A) Coupe PAS (en bas) et varves (NE) 7030-7180 (en haut, +2 cm). B) coupes de Comerford Dam (Vermont) de Lougee (1935b; COM de la fig. 8; en haut, +4 cm) et nouvelle coupe composite (COM à la fig. 8; au centre, +2 cm) et varves (NE) 7030-7160 de la vallée du Passumpsic (en bas, -1 cm). C) Coupes Barnett, BAR (en bas; au centre, +2 cm) et varves (NE) 7210-7350 (en haut, +4 cm).

Figs. 5, 8), match varve sections in the Passumpsic Valley (Antevs, 1922, site 86) spanning NE varves 7215-7350 (Fig. 10c). If the Littleton-Bethlehem Readvance moved east of the Comerford Dam, it either did not quite reach the Barnett section or readvance deposits occur just below the exposed part of the section between NE varves 7154 and 7215. The lower 40 couplets of the Barnett section appear to be ice-proximal couplets that are up to 20 cm thick and contain rippled sand beds with current directions of 190-220°. These couplets are lithologically similar to basal or ice-proximal varves at other localities in this area (Antevs, 1922, sites 85 and 86; Lougee, 1935b; base of section PAS on Figs 5, 8, 10a) and are difficult to match with the New England Varve Chronology. Upward in the Barnett section varves become progressively thinner, lose the sandy character present lower in the section, and are more easily matched to the New England Varve Chronology. Overall the Barnett section appears to be consistent with varve deposition in an ice-proximal environment followed by increasingly more distal meltwater deposition during ice recession.

Higher in the varve sequence near the Comerford Dam Antevs (1922) recorded a transition over about 20 yr time marked by the upward thickening of varves combined with a pronounced increase in clay content (NE varves 7330-7500, sites 85 and BAR, Figs. 5, 8, 10c). The top of the Barnett section contains this transition zone in which the increasing clay content of summer beds makes the distinction of annual couplets and varve measurement difficult. This clayey interval appears to be a product of ice recession. Ice recession in the Passumpsic and Moose River Valleys allowed the Moose River Valley to freely drain to the Passumpsic Valley instead of being forced to overflow into the Connecticut Valley by way of Chandler Brook and Miles Stream (Fig. 8). Recession of ice north of the threshold at Gilman in the Connecticut Valley caused the formation of Lake Coos that served as a settling basin. The decanting of water from one glacial lake into another such as occurred at Gilman has been recognized as a significant factor in reducing the volume and grain size of sediment supplied to a down valley lake (Smith, 1981; Smith and Ashley, 1985). Lakes down valley are also likely to have lower

underflow current velocities, which may allow more clay deposition. Ice recession in the Moose River Valley and in Lake Coos likely reduced the amount of silt and fine sand delivered to the Connecticut Valley immediately south of Lake Coos and may account for the clayey couplets beginning at NE varves 7330-7350.

PALEOMAGNETIC INVESTIGATIONS OF VARVES

PIONEERS IN SEDIMENTARY PALEOMAGNETISM

During the 1930's the New England Varve Chronology attracted the attention of geophysicists trying to document the secular variation of the geomagnetic field by studying sedimentary deposits (McNish and Johnson, 1938). The varve chronology provided an unparalleled time scale and the varves were composed of fine-grained sediment that was known to carry stable magnetic records. Using the varves in the Connecticut Valley Johnson *et al.* (1948) assembled a natural remanent magnetization record of declination (Fig. 11) for parts of both the lower and upper Connecticut Valley sequences of Antevs (1922, 1928). Johnson *et al.* (1948) took advantage of sites with long varve records, especially Antevs' (1922) Newbury section (site 73, Fig. 8), and at every section found an exact match between their varve measurements and Antevs' measured chronology. They did not create a complete inclination record because the results of laboratory sedimentation experiments reduced their confidence in the ability of sample inclination to faithfully record the geomagnetic field. In addition, their remanence results were never subjected to alternating field demagnetization (Zijderveld, 1967) in order to test the stability of samples. The removal of unstable components of magnetism with this technique is required in order to isolate detrital remanent magnetization as a record of the geomagnetic field.

REFINEMENT OF CONNECTICUT VALLEY PALEOMAGNETIC RECORDS

After the development of alternating field demagnetization, and statistical techniques for evaluating the precision of remanence data in the 1960's, Kenneth Verosub (1979a, 1979b) refined the work of Johnson *et al.* (1948) by formulating new detrital remanence records of declination and inclination for NE varves 3150-5500 (Figs. 11, 12). Using new exposures, and some originally studied by Antevs (1922), Verosub was able to match his varve records with the New England Varve Chronology. The first author and his students at Tufts University have been able to continue this work by collecting samples from varves at Canoe Brook and several exposures in the Connecticut, Passumpsic, and Merrimack Valleys (Figs. 11, 12, Appendix). At Newbury our reformulated declination and inclination records span NE varves 6963-8467. Paleomagnetic records for the entire New England Varve Chronology from NE varve 3150-8467 have now been reformulated except for results from NE varves 6601-6944. However, this interval partly overlaps the lower Connecticut Valley sequence leaving only NE varves 6850-6944 not covered by new results. Results obtained by Verosub (1979a, 1979b) and our new data are very similar to the results of Johnson *et al.* (1948). It

is important to note that the magnetic records from the lower and upper Connecticut Valley are compatible with the overlapping correlation of the upper and lower Connecticut Valley varve sequences (Fig. 7).

REGIONAL PALEOMAGNETIC CORRELATIONS

Remanent declination and inclination results provide useful records for testing interregional chronologic correlations (Brennan *et al.*, 1984; Ridge *et al.*, 1990, 1991, 1995; Pair *et al.*, 1994; Ridge, 1997), even with varve or other bedded mud sequences where a varve correlation is not possible. For the purposes of correlation declination records appear to be more useful because they are a more faithful record of the geomagnetic field and declination tends to vary over a range of about 70° . Inclination can be subject to flattening by sediment compaction and other depositional processes (see review in Ridge *et al.*, 1990), and geomagnetic inclination generally varies over a range of only 40° .

The remanent declination record for lake sediment in the Connecticut and Merrimack Valleys can be correlated with a compilation of declination records from New York (Fig. 11). The lake sediments used to construct the New York record are tied to glacial events that can then be magnetically correlated to varve sequences in New England, allowing a regional correlation of glacial events across several drainage basins. The most striking feature on the paleomagnetic declination correlation is a prominent westward swing in declination to $35\text{--}60^\circ$ West ($325\text{--}300^\circ$). This event is unique on both declination records and is a feature that can be used as a reference interval for regional correlation.

The inclination records from New England and New York can only be matched for the time period representing the last 1900 varve yr of the paleomagnetic record in New England (Fig. 12). Paleomagnetic records from prior to this time in New York come from sections in the western Mohawk Valley (Ridge *et al.*, 1990) that have flattened inclinations which negates their use for stratigraphic correlation. The Mohawk Valley varve sections occur beneath a great thickness of overlying sediment (15 m or more) and in most cases were compacted by overriding ice. The remanent inclination results from New England do not appear to suffer the same problems and are a more faithful record of geomagnetic inclination than records from the western Mohawk Valley of New York.

RADIOCARBON CALIBRATION OF VARVES

CALIBRATION OF THE LOWER CONNECTICUT VALLEY VARVE SEQUENCE

Organic sediment recovered from varves at Canoe Brook, Vermont (Fig. 1; Ridge and Larsen, 1990) has provided the only ^{14}C calibration of the lower Connecticut and Merrimack Valley varve sequences of Antevs (1922). A cluster of three ^{14}C ages on plant macrofossils in NE varve 6150 provide a calibration of 12.3^{14}C ka (Fig. 13; Table I). However, there may have been some lag between when the plants died and lacustrine deposition in which case the ^{14}C ages represent a maximum ("oldest possible") age for NE varve 6150. AMS ^{14}C ages of terrestrial macrofossils from varves in Sweden have a

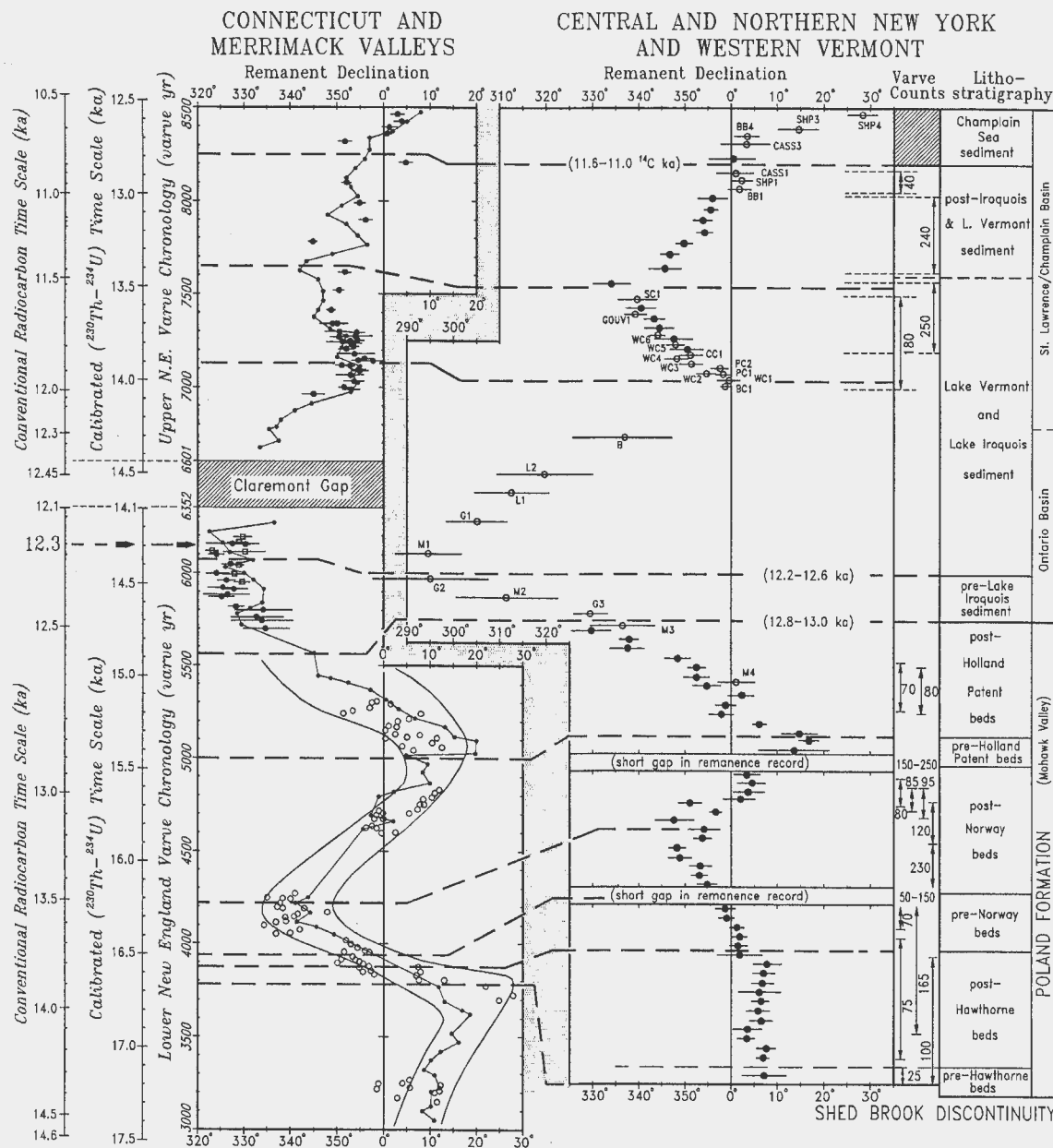


FIGURE 11. Correlation of Late Wisconsinan paleomagnetic declination records from New England and New York. All data except from Johnson *et al.*, (1948) are alternating field demagnetized at 17.5-40 mT and error bars are α_{95} (precision parameter) values. The Connecticut and Merrimack Valley declination record is plotted using the New England Varve Chronology as a time scale. The New York and western Vermont records are plotted by age relative to lithostratigraphic units, superposition at individual exposures, and geomorphic relationships (Brennan *et al.*, 1984; Ridge *et al.*, 1990; Pair *et al.*, 1994). Data in New England are from Johnson *et al.* (1948; dots with tie line), Verosub (1979a; open circles and envelope), and new results (Appendix) in the Merrimack Valley (open squares) and Connecticut Valley (dark circles). Data in New York are from Ridge *et al.* (1990; dark circles, Mohawk Valley), Brennan *et al.* (1984; open circles, eastern Ontario Basin), and Pair *et al.* (1994; open circles, St. Lawrence Basin), and new results in the Champlain Valley (dark circles; Appendix). The age for the beginning of Lake Iroquois is from Muller and Prest (1985), Muller *et al.* (1986), and Muller and Calkin (1993). The age of the Champlain Sea invasion shown on the New York record is based on Anderson (1988) and Rodrigues (1988, 1992).

Corrélations entre les relevés de déclinaison paléomagnétique de la Nouvelle-Angleterre et de l'état de New York. Toutes les données, sauf celles de Johnson *et al.*, sont démagnétisées (champ magnétique alternatif) à 17,5-40 mT et les traits horizontaux montrent un intervalle de confiance à α_{95} . La New England Varve Chronology a servi d'échelle temporelle au relevé de déclinaison des vallées du Connecticut et du Merrimack. Les relevés de l'état de New York sont reportés chronologiquement d'après les unités lithostratigraphiques, les superpositions à certains sites et les liens géomorphologiques (Brennan *et al.*, 1984 ; Ridge *et al.*, 1990 ; Pair *et al.*, 1994). Les données sur la Nouvelle-Angleterre sont de Johnson *et al.* (1948 ; points reliés entre eux), Verosub (1979a ; cercles non tramés et bordés) ; les nouvelles données (Appendix) concernent la vallée du Merrimack (carrés non tramés) et la vallée du Connecticut (cercles noirs). Les données de l'état de New York sont de Ridge *et al.* (1990 ; cercles noirs, vallée du Mohawk), Brennan *et al.* (1984 ; cercles non tramés, bassin de l'ouest de l'Ontario), et de Pair *et al.* (1994 ; cercles non tramés, bassin du Saint-Laurent) ; les nouveaux résultats concernent la vallée du lac Champlain (cercles noirs ; Appendix). La datation touchant le début du Lac Iroquois est de Muller et Prest (1985), Muller *et al.* (1986) et Muller et Calkin (1993). Le début de la Mer de Champlain a été daté par Anderson (1988) et Rodrigues (1988, 1992).

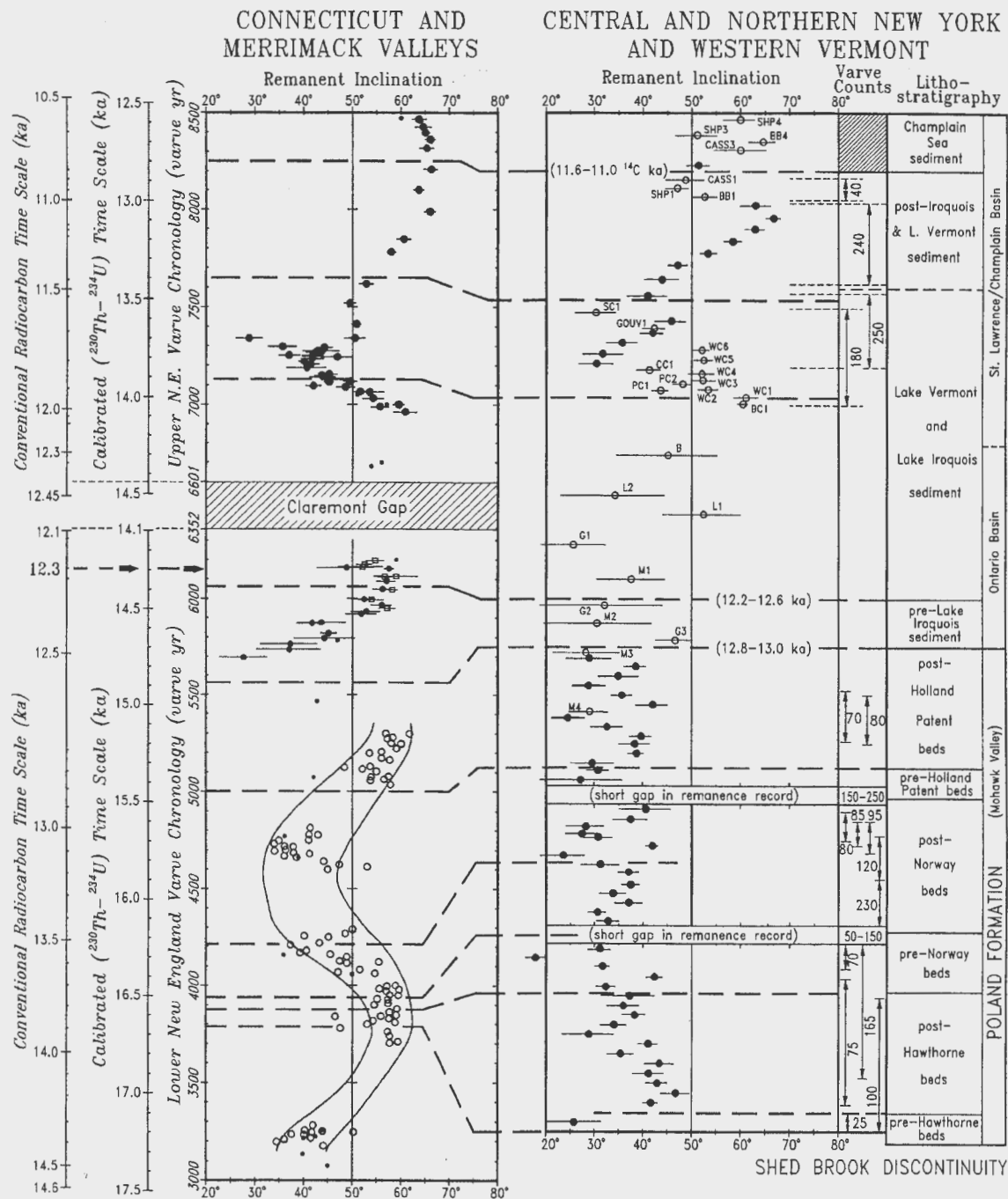


FIGURE 12. Correlation of Late Wisconsinan paleomagnetic inclination records from New England and New York. Correlation tie lines are based on declination results (Fig. 11). Data sources, plotting convention, and symbols are the same as for Figure 11.

Corrélation entre les relevés d'inclinaison paléomagnétique de la Nouvelle-Angleterre et de l'état de New York. Les lignes de lien corrélatif sont fondées sur les résultats de déclinaison (fig. 11). Se reporter à la figure 11 pour les sources des données, la méthode employée et les symboles utilisés.

scatter that reflects lags in deposition and the youngest ages appear to yield the most accurate estimate of true ^{14}C ages for the varves (Wohlfarth *et al.*, 1995). In previous publications (Ridge *et al.*, 1995, 1996; Ridge, 1997), the calibration of NE varve 6150 was recorded as 12.4 ^{14}C ka, the approximate average of the Canoe Brook ^{14}C ages. However, two of the ^{14}C ages at 12.35 ^{14}C ka are younger than 12.4 ^{14}C ka and

there was likely some lag in deposition, even if for only a few years. Plant macrofossils from varve 6150 have been analyzed by species and did not contain any aquatic vascular plant remains (N. Miller, pers. comm.). The macrofossils do include a small percentage of wet-soil sedges, the only macrofossils identified that could potentially cause ^{14}C ages to be too old. The plant types indicated that they were carried into

Lake Hitchcock from an open tundra-like environment (Miller, 1995). A full calibration of the lower Connecticut Valley sequence to both ^{14}C and calibrated (U-Th or inferred calendar) years (Fig. 13) is inferred using the CALIB 4.0 computer program (Stuiver and Reimer, 1993; Stuiver *et al.*, 1998). This calibration program accounts for disparities between the ^{14}C and calibrated ages resulting from the secular variation of atmospheric ^{14}C . Several compressions of the ^{14}C time scale occur from 12.4 to 10.7 ^{14}C ka as well as a prominent ^{14}C plateau at 12.6-12.4 ^{14}C ka. Recognition of ^{14}C variations in these time spans is critical to formulating an accurate ^{14}C chronology for glacial events in New Hampshire and Vermont.

One additional Canoe Brook ^{14}C age of 12.9 ka was obtained on a bulk sample of silt and clay that contained peat and gyttja fragments from NE varve 6156 (Table I). The fragments are rip-up clasts composed of fine organic sediment and sieving a few of them did not yield any identifiable plant macrofossils. Instead of ruining the sample with further sieving the remaining fragments were submitted with their enclosing silt and clay as a bulk sample. This type of organic material is not considered reliable for determining an atmospheric ^{14}C

calibration of a lacustrine sequence as compared to terrestrial plant macrofossils because it is sediment that was eroded from an older organic pond deposit and was later redeposited in Lake Hitchcock as ripped-up fragments. It may also contain the remains of aquatic species, especially algae that do not obtain their carbon directly from the atmosphere. The Connecticut Valley varves are calcareous due to marble and calcareous phyllite in Vermont. In addition to aged water from glacial melting, baseflow, and older organic sediment (Abbot and Stafford, 1996), the bedrock may have provided a source of carbon that could create anomalous ^{14}C ages for freshwater bodies. Bulk samples of organic lacustrine sediment, especially gyttja, have consistently yielded anomalous ages as compared to terrestrial plant macrofossils in attempts to calibrate the Swedish Varve Chronology (Wohlfarth *et al.*, 1993) and other lacustrine records (Oeschger *et al.*, 1985; Andr  e *et al.*, 1986). The 12.9 ka ^{14}C age for varve 6156 is older than the three other ^{14}C ages from NE varve 6150 at Canoe Brook (Table I) by about 450-600 ^{14}C yr (about 1200 varve or calibrated yr, Fig. 13) and these other dates were determined on terrestrial macrofossils that are generally considered more reliable mate-

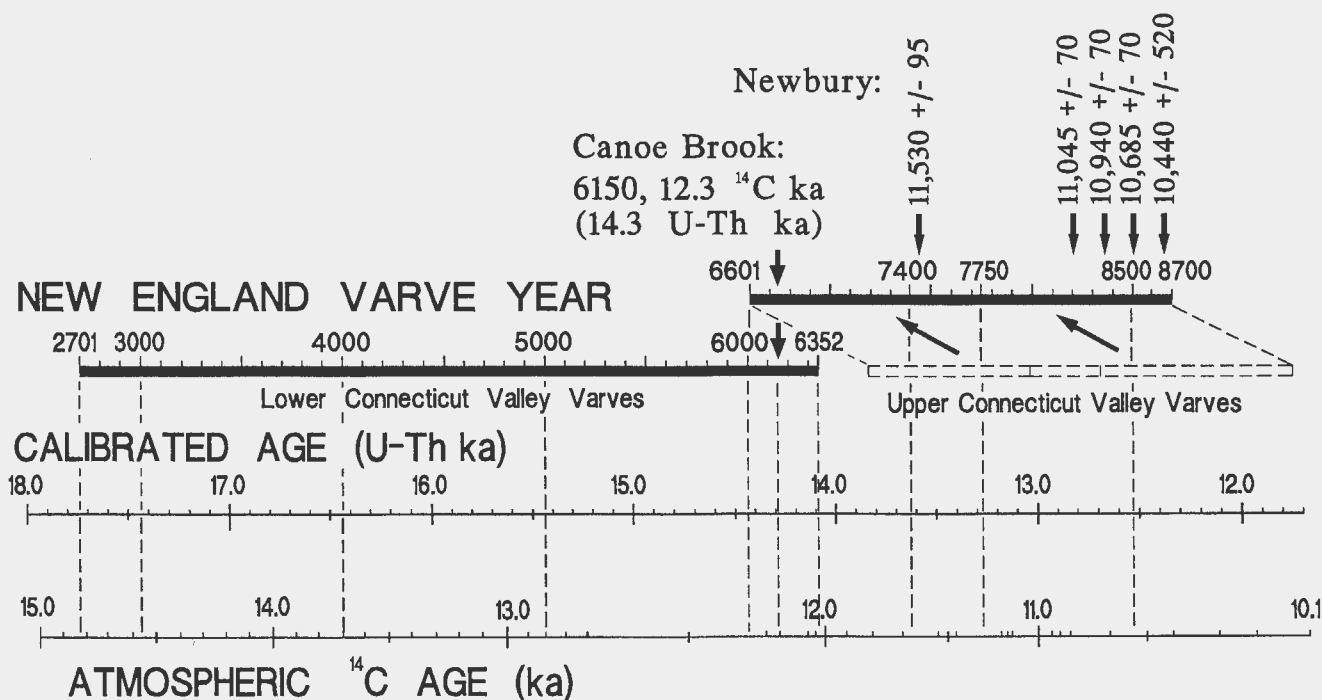


FIGURE 13. Calibration of the New England Varve Chronology. Time scales are based on a calibration point of 12.3 ^{14}C ka (14.3 cal. ka) for NE varve 6150 at Canoe Brook, Vermont (Table I). ^{14}C years were converted to calibrated (Th-U) years and the ^{14}C time scale was constructed using the CALIB 4.0 computer program (Stuiver and Reimer, 1993; Stuiver *et al.*, 1998). The CALIB program accounts for the disparity between ^{14}C and calibrated or calendar ages resulting from the secular variation of atmospheric ^{14}C . In the time span from 12.6 to 10.7 ^{14}C ka the ^{14}C time scale is especially non-linear with significant compressions of the ^{14}C time scale from 12.4 to 10.7 ^{14}C ka and a ^{14}C plateau at 12.6-12.4 ^{14}C ka. Four AMS and one conventional ^{14}C age from varves at Newbury, Vermont (Table I) are plotted for comparison.

  talonnage de la New England Varve Chronology. Les   chelles temporelles s'appuient sur le point d'  talonnage de 12,3 ^{14}C (14,3 ka   tal.) de la varve (NE) 6150 de Canoe Brook, au Vermont (tabl. I). Les ann  es au radiocarbone ont   t   converties en ann  es   talonn  es (Th-U) et l'  chelle temporelle au ^{14}C a   t     tablie avec le programme CALIB 4,0 (Stuiver et Reimer, 1993; Stuiver *et al.*, 1998). Le programme CALIB tient compte des divergences entre les datations au ^{14}C et les datations   talonn  es r  sultant des variations s  culaires du ^{14}C atmosph  rique. Au cours de la p  riode de 12,6    10,7 ^{14}C ka, l'  chelle temporelle en ^{14}C est particuli  rement non lin  aire avec des compressions importantes de 1,4    10,7 ^{14}C ka et un plateau de 12,6    12,4 ^{14}C ka. Quatre datations par spectrom  trie de masse et une datation au ^{14}C conventionnelle sur des varves    Newbury, au Vermont (tabl. I) sont montr  es pour fins de comparaison.

TABLE I
 ^{14}C ages from varves in the Connecticut Valley (locations on Fig. 1)

Laboratory number	Age (^{14}C yr BP)	$\delta^{13}\text{C}$ (‰)	NE varve number	Material dated	Reference
1. Canoe Brook, Vermont (Ridge and Larsen, 1990)					
GX-14231	12,355 \pm 75	-27.2	6150	Bulk silt and clay with, non-aquatic twigs and leaves	Ridge and Larsen, 1990
GX-14780	12,455 \pm 360	-27.6	6150	Handpicked non-aquatic leaves and twigs, mostly <i>Dryas</i> and <i>Salix</i>	Ridge and Larsen, 1990
CAMS-2667	12,350 \pm 90	—	6150	one <i>Salix</i> twig (AMS)	Norton Miller, pers. comm.
GX-14781	12,915 \pm 175	-27.1	6156	Peat and gyttja fragments	Ridge and Larsen, 1990
2. Newbury, Vermont (site 73 of Antevs, 1922)					
GX-23765	11,530 \pm 95	-27.0	7435-7452	Woody twig (AMS)	New
GX-23766	11,045 \pm 70	-27.5	8206	Woody twig (AMS)	New
GX-23640	10,940 \pm 70	-26.8	8357	Woody twig (AMS)	New
GX-23641	10,080 \pm 580	-26.7	8498-8500	Woody twig	New
GX-23767	10,685 \pm 70	-26.3	8504	Woody twig (AMS)	New
GX-23642	10,040 \pm 230	-26.5	8542-8544	Chunk of wood	New
GX-23643	10,440 \pm 520	-26.8	8652-8662	2 woody twigs	New
3. Columbia Bridge, Vermont (Miller and Thompson, 1979)					
WIS-961	11,540 \pm 110	-29.0	unknown (>7400)	Wood fragments	Miller and Thompson, 1979
WIS-919	11,390 \pm 115	-27.5	unknown (>7400)	Wood fragments	Miller and Thompson, 1979
WIS-925	20,500 \pm 250	—	unknown (>7400)	<i>Potamogeton</i> leaves and other plant remains	Miller and Thompson, 1979

Note: Blank spaces with hyphens indicate that information was not available from published source or was not obtained.

rials for ^{14}C dating. The 12.9 ^{14}C -ka age is also inconsistent with new ^{14}C ages from Newbury (discussed below) and paleomagnetic correlations between New England and New York.

The Canoe Brook calibration point (NE varve 6150 = 12.3 ^{14}C ka) can be tested using the paleomagnetic correlation between New England and New York. The Canoe Brook ^{14}C ages come from sediment that records the strong westward swing in declination that is a prominent part of both records (Fig. 11). Correlative sediment in New York represents the initiation of Lake Iroquois in the Ontario Basin at 12.6-12.2 ^{14}C ka (Muller and Prest, 1985; Muller *et al.*, 1986; Muller and Calkin, 1993) and has the same ^{14}C age as the sediment at Canoe Brook.

CALIBRATION OF THE UPPER CONNECTICUT VALLEY VARVE SEQUENCE

Until recently ^{14}C calibration of the upper Connecticut Valley varve sequence could only be inferred using three pieces of information: 1) the ^{14}C ages at Canoe Brook, 2) the correlation of the lower and upper Connecticut Valley sequences (Fig. 7), and 3) the CALIB program (Stuiver and Reimer, 1993). Four AMS ^{14}C ages and three conventional ^{14}C ages for woody twigs and wood from varves at Newbury, Vermont (Table I, Fig. 8; site 73 of Antevs, 1922) now provide a direct calibration of the upper Connecticut Valley varve sequence.

The new AMS ^{14}C ages (Table I) are consistent with the ^{14}C calibration of the New England Varve Chronology (Figs. 13, 14) formulated from the Canoe Brook ^{14}C ages and the CALIB 4.0 program (Stuiver and Reimer, 1993; Stuiver *et al.*, 1998). Because of their large precision parameters the conventional ^{14}C ages are not as useful for precisely calibrating the varve chronology. It is important to emphasize again that it has not been possible to fully evaluate lags related to the erosion and redeposition of terrestrial plant macrofossils in the lake. A lag is suggested by some of the Newbury ^{14}C ages that are younger than the inferred ^{14}C age of the varve sequence based on the Canoe Brook ^{14}C data (Figs. 13, 14). Two AMS ^{14}C ages from Newbury are about 350 ^{14}C yr older than the ^{14}C calibration (Fig. 14) but this difference occurs at a time when there is compression of the ^{14}C time scale and 350 ^{14}C yr corresponds to about 200 varve or calibrated years.

CALIBRATED VARVE AND PALEOMAGNETIC CHRONOLOGIES: 13-10.5 ^{14}C KA

DEGLACIAL EVENTS

A model of deglaciation for New Hampshire, Vermont, and adjacent New York (Fig. 15) is given here based on the varve chronology (Figs. 2, 4), the ages of basal varve localities (Figs. 5, 6), paleomagnetic correlations (Figs. 11, 12),

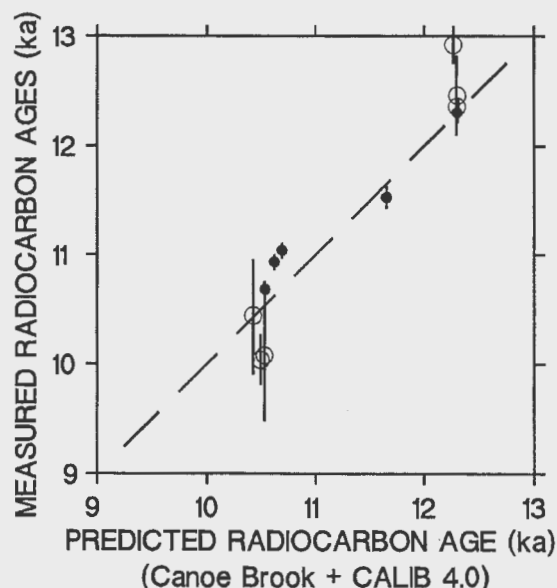


FIGURE 14. Plot of measured ^{14}C ages versus their predicted ^{14}C ages. Predicted ^{14}C ages are based on the positions of samples in the New England Varve Chronology and the ^{14}C calibration of the varves based on data at Canoe Brook (Fig. 13, Table I) and the CALIB 4.0 computer program (Stuiver and Reimer, 1993; Stuiver *et al.*, 1998). Open circles are conventional ^{14}C ages and smaller solid points are AMS ^{14}C ages. Error bars are ^{14}C age precision reported at $\pm 1\sigma$. The ages from Canoe Brook are the data points prior to 12.0 ka. Points above the calibration line represent measured ^{14}C ages older than the ^{14}C calibration defined by the Canoe Brook data and points below represent measured ^{14}C ages younger than the Canoe Brook calibration.

Graphique des datations au ^{14}C mesurées par rapport aux datations escomptées. Les datations escomptées sont fondées sur l'emplacement des échantillons dans la New England Varve Chronology et l'étalonnage des varves à partir des données du site de Canoe Brook (fig. 13, tabl. I) ainsi que le programme CALIB 4.0 (Stuiver et Reimer, 1993; Stuiver *et al.*, 1998). Les cercles non tracés représentent les datations au ^{14}C conventionnelles et les points noirs, les datations par spectrométrie de masse. Les traits verticaux donnent la précision des datations au ^{14}C reportée à $\pm 1\sigma$. Les points avant 12,0 ka représentent les datations de Canoe Brook. Les points situés au-dessus de la ligne d'étalonnage représentent les datations au ^{14}C plus vieilles que les datations étalonnées à partir des données de Canoe Brook et les points sous la ligne d'étalonnage représentent des dates mesurées plus jeunes que les dates étalonnées de Canoe Brook.

and the ^{14}C calibration of the varves (Fig. 13). According to these combined data sets deglaciation of the Connecticut Valley of New Hampshire and Vermont occurred between 12.6 and 11.5 ^{14}C ka (NE varve yr 5200-6000 plus 6612-7500). Disparities between varve and ^{14}C years during this period are a result of erratic changes in the atmospheric concentration of ^{14}C that have now been incorporated into a calibration of ^{14}C to calibrated or calendar years (Stuiver and Reimer, 1993; Stuiver *et al.*, 1998). Deglaciation of the Connecticut Valley (Figs. 5, 15) from Massachusetts to near Claremont, New Hampshire (12.6-12.45 ^{14}C ka, NE varve yr 5200-6000) occurred at a rate of 83.1 m/varve yr followed by a rate of 252 m/varve yr from Claremont to Littleton (12.45-12.0 ^{14}C ka, NE varve yr 6601-7000). Ice recession along a 30 km stretch of the central Merrimack Valley

occurred at 12.5-12.45 ^{14}C ka (NE varve yr 5700-6000) at a rate of about 100 m/varve yr (Figs. 6, 15). The Littleton-Bethlehem Readvance reached its maximum extent at 11.9-11.8 ^{14}C ka at which time the ice front was just north of the White Mountains. The drainage and flood events at NE varves 7200-7213 (Figs. 9a, 9b), suggested earlier to possibly represent the initial lowering of Lake Hitchcock, occurred at about 11.8 ^{14}C ka. Later drainage events appear to have occurred at 11.6 ^{14}C ka (NE varve 7500, Fig. 9c) and 11.1 ^{14}C ka (NE varve 7923, Fig. 9d). Following the Littleton-Bethlehem Readvance, ice recession from Littleton to the Québec border was completed by about 11.5 ^{14}C ka to allow the burial of plant fossils in varves near Columbia Bridge (Miller and Thompson, 1979). The minimum rate of deglaciation for this 70-km stretch of the valley from 11.8 to 11.6 ^{14}C ka (NE varve yr 7200-7500) would have been about 230 m/varve yr. Also, 11.6-11.5 ^{14}C ka approximately represents the time that lakes in the upper Connecticut Valley became entirely non-glacial as ice receded into Québec.

It has been possible to infer the approximate ^{14}C age of deglaciation in two areas of northern Vermont west of the Connecticut Valley (Fig. 15). Near Montpelier Antevs (1928) found basal varves (about NE varve 7050) from glacial Lake Winooski overlying till indicating deglaciation from this area at about 12.0 ^{14}C ka. In northern Vermont at Enosburg Falls in the Missisquoi Valley a section of 300 varves from large glacial lakes in the Champlain Valley (Lakes Fort Ann and Cándona of Parent and Occhiotti, 1988; Lakes Fort Ann and St. Lawrence of Pair and Rodrigues, 1993) was found beneath glaciomarine mud of the Champlain Sea. The basal varve at this section was not exposed and varves in the exposed section were thin (<1 cm). The paleomagnetic stratigraphy of this section indicates that it is correlative with varves in the Connecticut Valley having a ^{14}C age of 11.3-10.7 ^{14}C ka. The thin varves exposed at Enosburg Falls are probably a hundred to a few hundred years above the bottom of the varve stratigraphy, which would give the basal varves and deglaciation at Enosburg Falls an estimated age of 11.7-11.4 ^{14}C ka.

In the Hudson and Champlain Valleys of New York two glacial readvances have been recognized that may be equivalent to events in the Connecticut Valley (Fig. 15). In the Hudson Valley the Luzerne Readvance (Connally and Sirkin, 1971, 1973) occurred prior to 12.5 ^{14}C ka to allow the deposition of varves that were measured by Gerard De Geer and matched to varves in the Connecticut Valley (Antevs, 1922; NE varves 5501-5800, Figs. 2, 4). This readvance has been assigned to the Pt. Huron Stadial of the eastern Great Lakes region (Connally and Sirkin, 1973). No readvance has been recognized in New England from this time. However, large ice-contact deltas and subaqueous fans that were deposited in Lake Hitchcock in southern most Vermont (Larsen and Koteff, 1988) at about 12.6-12.5 ^{14}C ka, may represent an equivalent ice front position. Further north in the Champlain Valley the Bridport Readvance (Connally and Sirkin, 1973) may be the equivalent of the Littleton-Bethlehem Readvance in the Connecticut Valley based on its geographic position.

POSTGLACIAL EVENTS

In the upper Connecticut Valley a lake persisted until at least 1780 yr after deglaciation. Varves at Newbury contain plant fossils dating from 10.7-10.45 ^{14}C ka that are in the 200 youngest varves at the exposure that are overlain by non-varved lacustrine sand. The lake represented by the youngest varves at Newbury appears to have persisted until at least 10.4 ^{14}C ka and may have been seen by the first humans to enter the region. Archaeologists working in this section of the Connecticut and Passumpsic Valleys between Lyme and St. Johnsbury should focus their search for evidence of the earliest humans along the shoreline of this lake. In sections of the valley to the south where lakes drained much sooner after deglaciation the earliest evidence of humans should be on stream terraces. In the northern Connecticut Valley near Newbury stream terraces postdate very young (latest Pleistocene) lacustrine sediment and might not be the locations of the earliest evidence of humans.

The inferred age of the transition from lacustrine to marine sediment, and thus the invasion of the Champlain Sea, at Enosburg Falls (Fig. 1) is 11.1-10.6 ^{14}C ka based on paleomagnetic correlations to varves in the Connecticut Valley (Fig. 11). The paleomagnetic data from this time in the Connecticut Valley are not very precise and do not allow a precise correlation. The lacustrine to marine transition at Enosburg Falls occurs at the same time as a paleomagnetic declination of 0° as it changed from a western to eastern direction. This paleomagnetic signature for the Champlain Sea invasion has also been found in the western St. Lawrence region (Pair *et al.*, 1994). Our estimated age of 11.1-10.6 ^{14}C ka overlaps the span of ^{14}C ages proposed for this event in the western St. Lawrence Lowland (11.6-11.0 ka). Ages from the St. Lawrence Valley are based on marine ^{14}C ages from deep water fossils (Rodrigues, 1988, 1992) and the ^{14}C time frame that has been applied to pollen stratigraphy in southern Ontario (Anderson, 1988).

DISCUSSION: AN ACCURATE TERRESTRIAL ^{14}C CHRONOLOGY

Until recently the terrestrial chronology of deglaciation and other late Pleistocene events in New England has relied on ^{14}C ages from lake-bottom bulk sediment samples. Bulk organic samples from the bottoms of lake cores avoid problems associated with marine samples such as marine reservoir variations (Mangerud, 1972; Hjort, 1973; Mangerud and Gullikson, 1975; Bard, 1988; Bard *et al.*, 1994; Birks *et al.*, 1996) and the influences of meltwater in the marine environment (Sutherland, 1986; Hillaire-Marcel, 1988; Rodrigues, 1992). However, lacustrine bulk sediment samples can also yield anomalous old ages for other reasons (Shotton, 1972; Oeschger *et al.*, 1985; Andrée *et al.*, 1986; Wohlfarth, 1996). Aquatic plants as well as other aquatic organisms, which are frequently a significant part of bulk organic samples from lake cores, acquire carbon from lake water in which the concentration of ^{14}C may be lower than the atmosphere. Dissolved carbon taken up by aquatic plants may be from the atmosphere and organic and inorganic sources, and deliv-

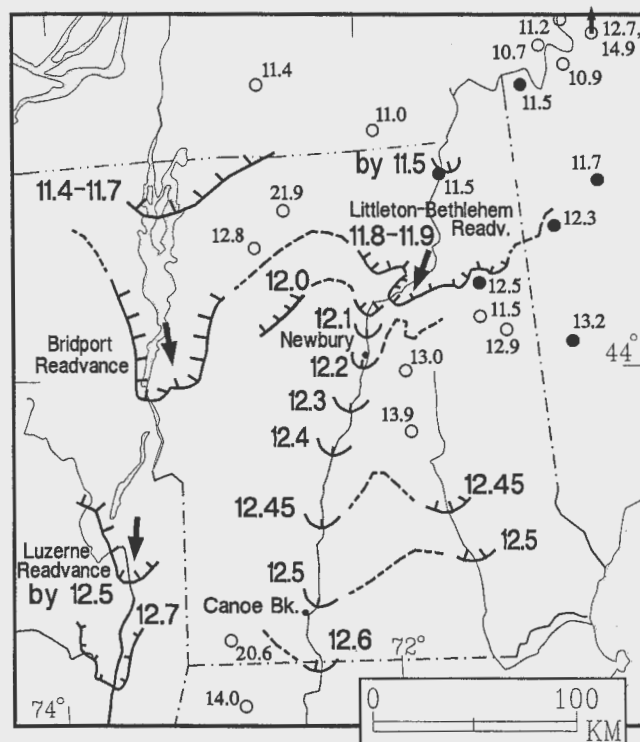


FIGURE 15. The deglaciation (dated ice front positions in ^{14}C ka) of New Hampshire, Vermont, and adjacent New York based on the combined ^{14}C calibration of the varve (Figs. 13, 14) and paleomagnetic declination chronologies (Fig. 11) and the varve ages for deglaciation in New England (Figs. 5, 6; see discussion in text). Arrows indicate glacial readvances. Both macrofossil (dark circles) and bulk sediment (open circles) lake-bottom ^{14}C ages (ka) relevant to the age of deglaciation are shown for comparison (Tables I and II). The apparently rapid deglaciation of southern New Hampshire and Vermont in ^{14}C years is an artifact of a ^{14}C plateau at 12.6-12.4 ^{14}C ka in which ^{14}C time changes very slowly as compared to varve or calibrated years (Fig. 13). Deglaciation of southern New Hampshire was actually slower than in northern New Hampshire, which is not apparent from the ^{14}C ages of ice front positions. As an example, deglaciation of southern New Hampshire at 12.6-12.4 ^{14}C ka represents 900 varve or calibrated yr while deglaciation of northern New Hampshire at 12.3-12.0 ^{14}C ka represents about 250 varve or calendar yr.

La déglaciation (fronts glaciaires en années ^{14}C ka) du New Hampshire, du Vermont et d'une partie de l'état de New York, reconstituée à partir de l'étalement combiné des chronologies établies pour les varves (fig. 13 et 14) et la déclinaison paléomagnétique (fig. 1) et de la déglaciation de la Nouvelle-Angleterre reconstituée à partir de l'âge des varves (fig. 5 et 6). Les flèches identifient les récessions. Les datations au ^{14}C (ka) de macrofossiles (cercles noirs) et de sédiments (cercles non tramés) de fonds lacustres relatives à la déglaciation sont données pour fins de comparaison (tabl. I et II). La rapidité de la déglaciation au sud du New Hampshire n'est qu'apparente en raison de l'existence d'un plateau de 12,6 à 12,4 ^{14}C ka durant lequel le temps au ^{14}C n'a évolué que très lentement en comparaison des années varvaires ou étalonnées (fig. 13). La déglaciation du sud du New Hampshire a en fait été plus lente qu'au nord, fait qui ne ressort pas de l'emplacement des fronts glaciaires. Par exemple, la déglaciation du sud du New Hampshire de 12,6 à 12,4 ^{14}C ka représentent 900 années varvaires ou étalonnées, tandis que la déglaciation du nord du New Hampshire de 12,3 à 12,0 ^{14}C ka représentent 250 années varvaires ou étalonnées.

ered to the lake by runoff or baseflow or generated internally (Abbott and Stafford, 1996). Dissolved inorganic carbon is derived from bedrock sources while organic carbon comes from the organic decay of lake bottom sediment, organic matter in soils, or older organic deposits. Particulate organic carbon and macrofossils that are older than the lake bottom at the time of deposition can also be delivered to a lake by runoff and shoreline erosion.

There is clear evidence in New England indicating that bulk sediment and aquatic plants can produce significant ^{14}C errors. Aquatic plants that have anomalous ^{14}C ages have been found in varves in the upper Connecticut Valley near Columbia Bridge (Table I). Macrofossil debris of the plant *Potamogeton* has a ^{14}C age of 20.5 ka while non-aquatic plant macrofossils from horizons above and below it have ^{14}C ages of 11.5–11.4 ka. Bulk organic sediment from the bottoms of ponds in Vermont has yielded ^{14}C ages of 20.6 and 21.9–19.6 ka (Table II, Fig. 15). Basal sediment from Lower Togue Pond adjacent to Mt. Katahdin in Maine has yielded a basal ^{14}C age of 21.3 ka (Table II). Basal bulk sediment from Unknown Pond, Maine has yielded ^{14}C ages 1500–3700 yr older than similar materials from lakes to the south (Table II, Fig. 15). Anomalous old ages for bulk sediment have also been recognized further north in New Brunswick (Karrow and Anderson, 1975). These bulk sediment ^{14}C ages have extreme errors (>1000 yr) that are easily recognized because they clearly do not fit the younger lake-bottom chronology from surrounding areas. However, comparing bulk ^{14}C ages with an existing lake bottom chronology that is also based on bulk sediment ages does not allow us to recognize more subtle errors (<1000 yr) that may be pervasive. This can only be accomplished with a direct comparison of bulk sediment and plant macrofossil ages, which have systematically been different by 300–800 yr in European lake sediment studies (Oeschger *et al.*, 1985; Andr  e *et al.*, 1986; Wohlfarth, 1996). Although it is possible for a bulk sediment sample to yield an accurate atmospheric ^{14}C age, systematic testing for subtle errors has not been incorporated into any chronology in New England. Until recently it has been customary to use the oldest lake bottom bulk sediment ages as minimum ("youngest possible") atmospheric ^{14}C ages for deglaciation but this ignores the errors that cause ^{14}C ages to be older than the actual atmospheric ^{14}C age of deglaciation. In general, lake bottom bulk sediment samples yield 'terrestrial ^{14}C ages' that do not meet the strict criteria necessary to consistently be used as atmospheric ^{14}C ages or to formulate an accurate glacial chronology.

One technique that should be applied to bulk sediment and macrofossil ^{14}C ages that can help to screen samples with anomalous ages is the analysis of ^{13}C . The measurement of $\delta^{13}\text{C}$ values represents not only a means of correcting for fractionation (Stuiver and Polach, 1977), but in some cases may indicate contamination or non-atmospheric sources of carbon. Non-aquatic plant leaves and wood, the best material for acquiring an atmospheric ^{14}C age, generally have $\delta^{13}\text{C}$ values of -28 to -26‰ (Stuiver and Polach, 1977; Lini *et al.*, 1995). $\delta^{13}\text{C}$ values outside of this range

indicate that the associated ^{14}C age may be too old because of aquatic plants or contamination by 'old' carbon. Unfortunately, $\delta^{13}\text{C}$ values within the non-aquatic plant range do not provide unequivocal proof that a ^{14}C age is free of potential errors and other factors such as sample type must be evaluated as well.

The most accurate ages for determining an atmospheric ^{14}C chronology are those from non-aquatic plant macrofossils, which avoid problems with marine and freshwater aquatic organisms. In order to avoid potential errors inherent to aquatic fossils, non-aquatic samples should not be mixed with aquatic organisms. A high precision can be obtained with AMS ^{14}C ages on relatively small samples that allow the dating of specific plant species. However, non-aquatic plant macrofossils still present a potential problem when trying to interpret results. All plant macrofossils are potentially susceptible to the problem of a lag in erosion and redeposition after the plant dies on the land surface. Preservation of macrofossils in a frozen or very cold soil or anoxic wetland sediment may allow the survival of this material for millennia before it is eroded and transported to a lacustrine setting. Woody materials, but also leaves, may survive destruction during erosion and transport to become macrofossils that will produce a ^{14}C age that is significantly older than the clastic sediment in which it is found. A systematic scatter among macrofossil ages in contemporaneous varves in Sweden appears to be the result of lags in deposition and ages on the younger side of this scatter are now accepted as being more accurate representations of the ^{14}C age of enclosing sediment (Wohlfarth *et al.*, 1995). Unless the magnitude of depositional lags can be evaluated macrofossils should be treated as yielding maximum ("oldest possible") atmospheric ^{14}C ages for the enclosing lacustrine sediment. If possible multiple ^{14}C ages from a single horizon, or spread across a known number of varves, should be used to test the contemporaneity of the macrofossils and their enclosing sediment.

Another possible problem that may exist for macrofossils deposited in shallow ponds and lakes that have high organic productivity and low sedimentation rates is the coating of macrofossils by aquatic algae. If the algae are not removed by laboratory sample treatment it will cause the ^{14}C age for the macrofossil to be too old. This should not be a problem in large glacial or non-glacial lakes that are deep and retard the penetration of light to the floor of the lake. Also, if a lake has a relatively high clastic sedimentation rate the burial of macrofossils will occur before significant growth of algae.

EVALUATION: CALIBRATED VARVE AND PALEOMAGNETIC CHRONOLOGY

As indicated earlier our ^{14}C -calibrated varve and paleomagnetic chronology for New England appears to be consistent with the ^{14}C chronology of the western St. Lawrence Lowland (Anderson, 1988; Rodrigues, 1988, 1992). However, the new chronology is not consistent with existing chronologies for western New England (Davis and Jacobson, 1985; Hughes *et al.*, 1985; Stone and Borns, 1986; Dyke and Prest, 1987) based on bulk sediment ^{14}C ages from cores of small lakes

TABLE II

 ^{14}C ages from lakes in northern New England and adjacent Québec that are relevant to the age of deglaciation

Location	Laboratory	Age (^{14}C)	$\delta^{13}\text{C}$ (‰)	Material dated	Reference (%)
A. Macrofossil samples from lake cores					
Pond of Safety, NH	OS-7125	12,450 \pm 60	-18.2	<i>Dryas</i> , moss parts, <i>Carex</i> seeds, <i>Daphnia</i> , insect parts <i>Salix herbacea</i> , <i>Characea</i>	Thompson <i>et al.</i> , 1996
Cushman Pond, Me	OS-7122	13,150 \pm 50	-23.5	<i>Salix herbacea</i> leaves plus insect parts, <i>Dryas</i> , moss parts, <i>Characeae</i> , <i>Daphnia</i> , and woody twigs.	Thompson <i>et al.</i> , 1996
Surplus Pond, Me	OS-7119	12,250 \pm 55	-28.1	<i>Salix herbacea</i> leaves	Thompson <i>et al.</i> , 1996
Spencer Pond, Me	AA-9506	11,665 \pm 85	-	(not reported)	C. Dorian, unpub. in Thompson <i>et al.</i> , 1996
Lower Black Pond, Me	OS-7123	11,500 \pm 50	-28.1	Woody twigs plus other macrofossils	Thompson <i>et al.</i> , 1996
B. Bulk sediment or bulk organic sediment samples from lake cores					
Hawley Bog Pond, Ma	WIS-1122	14,000 \pm 130	-	Basal organic silt	Bender <i>et al.</i> , 1981
Ritterbush Pond, Vt	CAMS-20197	21,860 \pm 370	-24.8	Bulk sediment, 6 cm above base of pond sediment.	Lini <i>et al.</i> , 1995; Paul Bierman, pers. comm.
	CAMS-32852	20,740 \pm 530	-25.8	Bulk sediment, 19 cm above	Lini <i>et al.</i> , 1995; Paul
	CAMS-33133	20,110 \pm 170	-25.8	base of pond sediment.	Bierman, pers. comm.
	CAMS-33349	19,570 \pm 170	-25.8	(3 replicates with additional acid wash)	
	CAMS-20902	11,940 \pm 90	-34.6	Gyttja, 50 cm above base of pond sediment.	Bierman <i>et al.</i> , 1997
Woodford Bog, Vt	GX-16951	20,575 \pm 1250	-	Bulk detrital organics	Davis <i>et al.</i> , 1995
Sterling Pond, Vt	CAMS-17895	12,760 \pm 70	-	Basal gyttja	Lin <i>et al.</i> , 1995
Mirror Lake, NH	BX-5429	13,870 \pm 560	-	Bulk sediment with <i>Dryas</i>	Davis <i>et al.</i> , 1980; Davis and Ford, 1982
Deer Lake Bog, NH	QL-1133	13,000 \pm 400	-	Bulk sediment	Spear, 1989
Lost Pond, NH	QL-985	12,870 \pm 370	-	Bulk sediment	Davis <i>et al.</i> , 1980
Lake of the Clouds, NH	I-10684	11,530 \pm 420	-	Basal bulk sediment	Spear, 1989
Lower Togue Pd., Me (not on Fig. 14)	SI-2732	21,300 \pm 1900	-	Bulk laminated sediment, basal 10 cm pond sediment	Davis and Davis, 1980
	SI-2992	11,630 \pm 260	-	Bulk laminated sediment, 50 cm above base	
Unknown Pond, Me (2 ages)	GSC-1339	14,900 \pm 240	-	Basal organic sediment	Mott, 1981
	GSC-1404	12,700 \pm 280	-	Basal organic sediment above first sample	
Boundary Pond, Me	GSC-1248	11,200 \pm 200	-	Basal organic sediment	Mott, 1981
Chain of Ponds, Me	-	10,860 \pm 160	-	Bulk sediment	Borns and Calkin, 1977
Mount Shefford, Qc	-	11,400 \pm 340	-	Basal bulk sediment	Richard, 1977 (as reported in Davis and Jacobson, 1985)
Barnston Lake, Qc	GSC-420	11,020 \pm 330	-	Bulk organic sediment	McDonald, 1968
Lac aux Araignées, Qc	GSC-1353	10,700 \pm 310	-	Basal organic sediment	Mott, 1981

Notes: Locations listed by state/country and are shown on Figure 15. List partly compiled from Thompson *et al.*, 1996.

This list does not include ages from varve sections in the Connecticut Valley that are listed on Table I.

Blank spaces with hyphens indicate that information was not available from published source or was not obtained.

(Table II). Our atmospheric ^{14}C chronology indicates that deglaciation of New Hampshire and Vermont began at 12.6 ka in southern Vermont and ended with ice receding into Québec at about 11.5 ka (Fig. 15). The other models show deglaciation of this region completed before 13.5-13.0 ^{14}C ka. We interpret the bulk sediment ^{14}C ages used to construct the existing chronologies as maximum ages for deglaciation and not accurate representations of the atmospheric ^{14}C age of deglaciation for reasons discussed above. Deglaciation models formulated in the early 1980's used the only available data,

which included no AMS ^{14}C ages and no samples composed of only non-aquatic plant macrofossils.

In general our chronology appears to be consistent with recently obtained ages for deglaciation based on non-aquatic plant macrofossils across northern-most New England (Fig. 15, Table II). In particular the proposed age of the Littleton-Bethlehem Readvance (11.9-11.8 ^{14}C ka) appears to be consistent with macrofossil and bulk sediment ^{14}C ages (Thompson *et al.*, 1996, 1999) north of this feature. The only

^{14}C ages that are inconsistent with the proposed chronology are from bulk samples and samples of mixed assemblages of aquatic and non-aquatic macrofossils from south of the Littleton-Bethlehem Readvance. Bulk sediment ages just south of this ice margin indicate a time of deglaciation that is up to 1500 yr older than predicted by our chronology. In front of the Bethlehem Moraines there are three macrofossil ages (Cushman Pond, Pond of Safety, and Surplus Pond, Table II), only one of which is in agreement with our age for the Bethlehem Moraines (11.9–11.8 ^{14}C ka). The ^{14}C age from Cushman Pond (12.9 ka) appears to differ by about 1000 yr from our model but this age has a $\delta^{13}\text{C}$ value of -23.5‰ , somewhat lower than expected for non-aquatic plants that represent an accurate ^{14}C record. The ^{14}C age from Pond of Safety (12.5 ka) is about 400 yr older than expected from our chronology and has a suspect $\delta^{13}\text{C}$ value of -18.2‰ . These two ^{14}C ages are from samples that contain non-aquatic plants but also aquatic organisms such as *Daphnia* (aquatic crustacean), *Characea* (aquatic algae), *Carex* (sedge) seeds, and moss parts. It should not be assumed that these fossils provide accurate atmospheric ^{14}C ages (N. Miller, pers. comm.). Also included in the apparently anomalous samples are insect parts that will have a ^{14}C age reflecting the diet of the insect. Depending on the insect species the diet may or may not include aquatic plants or other organisms that feed on aquatic plants. The only macrofossil ^{14}C age along the Bethlehem Moraines that is in agreement with our chronology is from Surplus Pond ($12,250 \pm 55$, 12.3 ka on Fig. 15). This sample has a $\delta^{13}\text{C}$ value of -28.1‰ and is composed entirely of leaves from a non-aquatic scrub willow (*Salix herbacea*). It is our interpretation that many of the bulk sediment ages and some of the macrofossil ages may be in error because they were obtained on samples that did not exclusively contain fossils of non-aquatic plants. Only non-aquatic plant macrofossils (leaves and woody twigs) with $\delta^{13}\text{C}$ values of -26.3 to 27.6‰ have been used to calibrate the New England Varve Chronology (Fig. 13) and formulate our deglaciation chronology (Fig. 15).

CONCLUSIONS

Any formulation of a regional synthesis of deglaciation requires that events in different regions be compared with compatible time scales and that correlations be tested with as many techniques as possible. For the past 50 years a synthesis of deglacial events across New England has been plagued by a scarcity of numerical ages and inconsistencies between them. There was no alternative to accepting ^{14}C ages at face value from both marine and lake bottom environments and assuming they were roughly measurements of an atmospheric ^{14}C -time scale. An independent stratigraphic test of the contemporaneity of glacial events and accurate atmospheric ^{14}C ages have been needed, especially if accurate comparisons are to be made to climatic data from marine and ice core records. The New England Varve Chronology, despite its existence for more than 75 years, appears just now to be poised to fill this void, especially when combined with other lithostratigraphic data, paleomagnetic records, and AMS ^{14}C

ages from macrofossils of non-aquatic plants. The ^{14}C -calibrated varve and paleomagnetic records for New England and New York (Figs. 11–13) have here been used to compile a comprehensive atmospheric ^{14}C chronology of deglaciation (Fig. 15) and postglacial events in northwestern New England and surrounding areas. At the very least, the combination of these techniques has established an independent test of other deglacial chronologies and an alternative to chronologies based on ^{14}C ages of marine fossils, aquatic plants, or bulk organic samples from lakes. The combined varve, paleomagnetic, and atmospheric ^{14}C chronologies hold the promise in the coming years of providing a correlation tool of unparalleled resolution.

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Article B. SPILLWAY GEOLOGY

excerpted from Koteff, Carl, 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 archeology, in J. Brigham-Grette, ed., Field trip Guidebook, American Quaternary Association 1988: University of Massachusetts Department of Geology and Geography Contribution 63, p. 169-208.

INITIATION OF GLACIAL LAKE HITCHCOCK

The inception of Lake Hitchcock was dependent on the presence of an earlier and higher glacial lake, Lake Middletown (Stone et al., 1982) (fig. 3), in the Connecticut River valley at Middletown and the tributary Mattabesset River valley during retreat of the Connecticut Valley lobe of the Laurentide ice sheet. An extensive deltaic complex controlled by the level of Lake Middletown completely filled the Connecticut River valley in the vicinity of Rocky Hill. This sediment mass later formed the dam for Lake Hitchcock. Formation of such drift dams in south-draining valleys has been found to be a necessary condition for the creation of many glacial lakes in southern New England. Lake Middletown itself was impounded by a long mass of older meltwater sediments that effectively filled the lower Connecticut River valley southeast of Middletown. Because these deposits extended at least 12 mi (20 km) down the valley, entrenchment of them and consequent lowering of Lake Middletown was relatively slow.

Construction of the deltaic complex began with deposition of successive, contiguous ice-marginal deltas in Lake Middletown in the Cromwell area, which blocked that relatively narrow part of the Connecticut River valley (see STOP 1, fig.5). As the ice margin retreated from the Cromwell deltas, meltwater was impounded behind them at a very slightly higher level than Lake Middletown, and ice-marginal deltas formed in this higher lake near Rocky Hill and on the east side of the Connecticut River in Glastonbury. The waters of this relatively small lake spilled over the Cromwell deposits. A well-developed channel (see STOP 1, figs. 5 and 6), called the Dividend Brook Spillway (Hartshorn and Koteff, 1968) was carved into the Cromwell delta surface. Together, the Cromwell-Rocky Hill-Glastonbury deltas are referred to as the drift dam at Rocky Hill.

Ice retreat during the formation of the deltaic complex uncovered, west of the Connecticut River, a bedrock upland that now forms the east-west divide between two tributaries of the Connecticut River, the Mattabesset River to the south and the Park River to the north (fig. 3). The small lake controlled by the Dividend Brook spillway expanded northward behind the deltaic complex, east of Cedar Mountain. At the same time, Lake Middletown expanded northward from the Mattabesset basin across a low part of the divide in the New Britain-Newington area, west of Cedar Mountain. As ice retreated from the north end of Cedar Mountain, the lake level behind the delta complex dropped and the lake coalesced with Lake Middletown. The Dividend Brook spillway was abandoned and erosion of its channel ceased. The final floor altitude of this spillway was 129 ft, controlled by the level of Lake Middletown into which it drained. The deltaic complex therefore survived to constitute the dam for Lake Hitchcock.

Although Lake Middletown continued to lower slowly by entrenchment of its drift dam, it remained high enough to cover the low part of the divide (no higher than 110 ft (34 m)) in the New Britain-Newington area and the lake was able to expand northward into the Connecticut River basin during ice retreat. Altitudes of deltas on both the east and west side of the basin indicate that Lake Middletown persisted, but with slowly lowering levels, until the ice margin retreated as far north as Windsor.

Further lowering of Lake Middletown allowed emergence of the low divide area at New Britain-Newington, and separated the shrinking Middletown lake to the south and the first phase of glacial Lake Hitchcock to the north. As the ice margin in the Connecticut Valley retreated northward, Lake Hitchcock expanded in area although its level gradually lowered because of erosion of till, waterlaid sediments, and weathered bedrock in the spillway. Meltwater-fed deltas were constructed successively northward in the lake during stagnation-zone retreat, and their lowering altitudes northward reflect the erosion of the drift at the New Britain channel area. This period of lowering, referred to here as the Connecticut phase of Lake Hitchcock, lasted until the floor of the New Britain channel stabilized on resistant bedrock, preventing further lowering of the lake level, and initiating the stable phase of Lake Hitchcock. By this time, the ice margin may have been as far north as Chicopee, Mass., but its exact position is still unclear.

During its stable phase, Lake Hitchcock continued to expand as the ice margin retreated north from Chicopee through all of Massachusetts and much of New Hampshire and Vermont. Meltwater-fed deltas were built successively in the lake probably to about Burke, Vt. The stagnation-zone retreat of the margin was generally systematic, interrupted in places by local readvances such as one at Chicopee (Larsen, 1982). Most of these readvances have been identified only in recent years, and no doubt others will be found as new exposures become available. However, none of them is believed to represent more than local and short-lived events and thus are not correlated regionally.

Lake Hitchcock was once thought to have drained catastrophically when the ice margin had reached just north of Hanover, N.H. (Lougee, 1939, 1957). Recent work by Koteff and Larsen (1985, in prep.) on postglacial uplift studies, however, has established the longer lake to Burke; also, the presence of post-lake stream terraces along the Connecticut River at the Cromwell-Rocky Hill-Glastonbury drift dam only about 30 feet (10 m) below the projected level of Lake Hitchcock indicates a somewhat less dramatic end to the lake.

Although we still lack definitive evidence to place absolute ages on the inception and the demise of glacial Lake Hitchcock, its life-span can be placed in a regional time-frame. From a regional array of radiocarbon dates and correlated ice-marginal deposits, Stone and Borns (1986) have placed the ice margin in the southern part of the lake basin at 16,000-17,000 years B.P. Varve correlation studies (Antevs, 1922) indicate a minimum of 4000 years for the life of the lake. Koteff and Larsen (1985), emphasizing radiocarbon dates reported by Davis and Ford (1982) from the White Mountains area of New Hampshire, suggest that Lake Hitchcock was still in existence, with its level controlled by the New Britain spillway, at 14,000 years B.P. Larsen (1984) provided evidence for drainage of Lake Hitchcock before 12,500 years ago, and perhaps even before 13,000 years ago while the ice sheet remained in the Winooski River valley of northern Vermont. A radiocarbon date of 12,200 \pm 350 years B.P. was reported by Colton (1960)

from wood in a small peat bog excavated during runway construction at Bradley Airport, Windsor Locks, CT (see fig. 11). Previously, little significance as to the time of lake drainage was placed on the 12,200 year date. It now appears from our regional compilation that this dated locality lies on the lake-bottom surface in front of the Bradley Field delta, and that the date thus records a time after the lake had drained. A radiocarbon date of 10,710 \pm 330 years B.P. (Preston, Person, and Deevey, 1955) from transported wood fragments in gravel at the mouth of the New Britain spillway channel (fig. 7) was thought by Flint (1956) to represent a time when Lake Hitchcock still spilled through this channel. Persistence of the lake until this 10,710 year date seems unlikely in view of recent evidence; re-evaluation of the stratigraphic horizon from which this dated material was taken is still needed.

Article C. PINGO GEOLOGY

excerpted from Stone, J.R. and Ashley, G.M., 1992, Ice-wedge casts, pingo scars and the drainage of glacial Lake Hitchcock, Trip A-7 in Robinson, Peter, and Brady, J.B., eds., Guidebook for fieldtrips in the Connecticut Valley region of Massachusetts and adjacent States, New England Intercollegiate Geological Conference 84th Annual Meeting, Amherst, Mass., Oct. 9-11, 1992: University of Massachusetts, Geology and Geography Contribution 66, vol. 2, p. 305-331

PINGO SCARS ON DRAINED GLACIAL LAKE-BOTTOM SURFACES

Clusters of circular to subcircular shallow depressions, many with subtly raised rims, occur on drained glacial lake-bottom surfaces in Connecticut and Massachusetts (figs. 1 and 2). They are interpreted as pingo scars (Stone and Ashley, 1989; Stone and others, 1991) based on; 1) their striking resemblance in morphology and distribution to ramparted depressions identified as pingo scars and other ground-ice depressions in northwestern Europe (Svensson, 1969; Mitchell, 1973; De Gans, 1988), 2) the deformational character of their internal structure as seen in a cross-sectional exposure (fig. 5, Stop 3) and on ground penetrating radar (GPR) lines across the surface features, and 3) the age and character of the clastic and organic fill within the deformational structures.

The surface depressions generally have less than 10 ft (3 m) of relief and therefore do not show up well on 1:24,000-scale topographic maps with 10-ft contour interval. They were not reported or described on any of the detailed surficial geologic maps of the area, but were first noticed during compilation and synthesis of the detailed maps for the State Quaternary Geologic Maps of Connecticut (Stone and others, 1992; Stone and Schafer, in prep.). Comprehensive and detailed air-photo analysis has revealed these clustered depressions to be characteristic of former lake-bottom surfaces, the most extensive of which are those of glacial Lake Hitchcock.

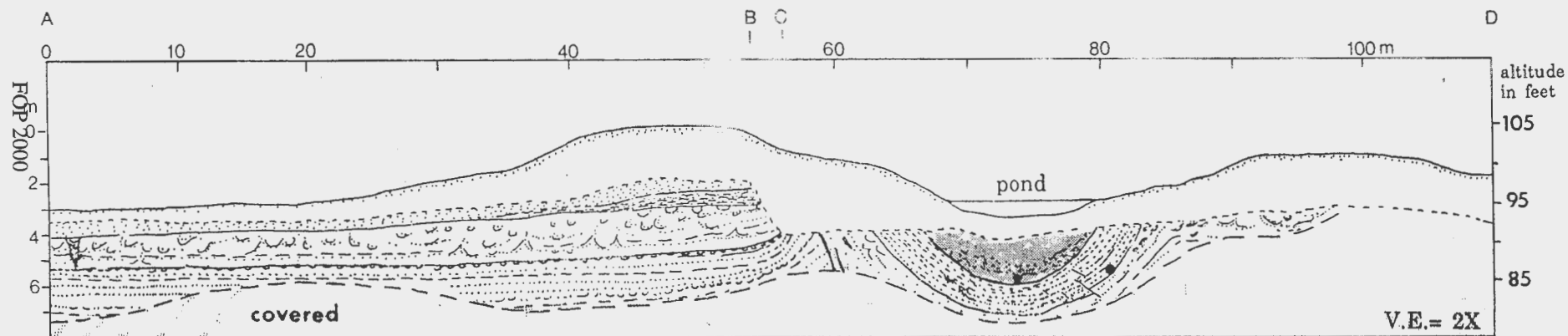
Morphology and Distribution

These landforms occur predominantly on surfaces underlain by fine-grained lake-bottom sediments (fine sands, silts and clays). Locally they are found developed on till and sandy spit-deposit surfaces on the sides of drumlins that were islands in Lake Hitchcock. They occur only below paleo-lake level. In places, they are closely associated with eolian deposits and dunes. They are not found on delta surfaces which were above paleo-lake level; they are absent on fluvially-modified post-lake surfaces such as stream terraces and floodplains. They occur in clusters of isolated and mutually interfering forms. Maximum densities of pingo scars are approximately 150/km². Diameters are generally from 20 to 40 m, but range from 5 to 100 m; one depression has a diameter of 250 m. The surface depressions are generally 1-3 m deep. Rims range from 0.3 to 1.5 m high. They are clearly visible on large scale maps with contour intervals of 1 and 2 ft, such as the Metropolitan District Commission (MDC) 1:2,400-scale maps (figs. 5B, 9B, and 12) and (fig. 12, Stop 6) and on 1:12,000-scale air photos. Closed depressions that contain small ponds, or wetlands/vernal pools are common, as well as many features in which the rims are breached and C-shaped contours mark the pingo scars.

Internal Structure

Stratigraphy and structure beneath the surface depressions are known from a roadcut exposure on Route I-91 between exits 34 and 35 (fig. 5 and Stop 3). Ground penetrating radar (GPR) data supplemented by vibracores across several other depressions add to our knowledge of the internal structure. The following internal characteristics (fig. 5) have been observed:

- 1) Beneath the outer edges and rim of the surface depression, upper lacustrine fine-sand and silt beds are gently deformed upward as much as 1.5 m and are broken by fractures. Inward from the fracture zone, lacustrine beds are intensely collapsed as much as 4 m downward. Beneath the central part of the depression, lacustrine materials are completely homogenized as a massive gray silt.
- 2) The structural depression created by collapse of the lacustrine beds is filled with a bowl-shaped body of medium sand in concentrically dipping beds. These beds display small-scale reverse faults displacing them downward toward the center of the structural depression and indicate that filling was synchronous with later stages of collapse of the underlying lacustrine materials. The sand is coarser grained than most of the lacustrine sediment. This fact and the nature of the bedding indicates that the sand fill is probably eolian-supplied material that settled out in ponded water in the collapsing depression.
- 3) Further filling of the structural depression consists of a 2-m-thick pod of gyttja and peat above the sand and below the center of the surface depression. Sandy clastic lenses extend into the peat body from the sides.



A

- Original topography (from MDC map 91)
- Top of section as exposed, August 1989
- * ^{14}C date 11,890+130
- ^{14}C date 14,330+430
- Glaciolacustrine beds; silts, clays and f-vf sands
 - ← zone of pervasive water-escape convolutions
 - ← marker bed of sand ripples with silt/clay drape
 - ← non-convoluted zone
- Depression-fill sand beds; probably eolian-supplied material
- Depression-fill organic material
- Fine-sand and silty-sand beds unconformably overlying convoluted lacustrine beds; probably remnant of "rim" material

B

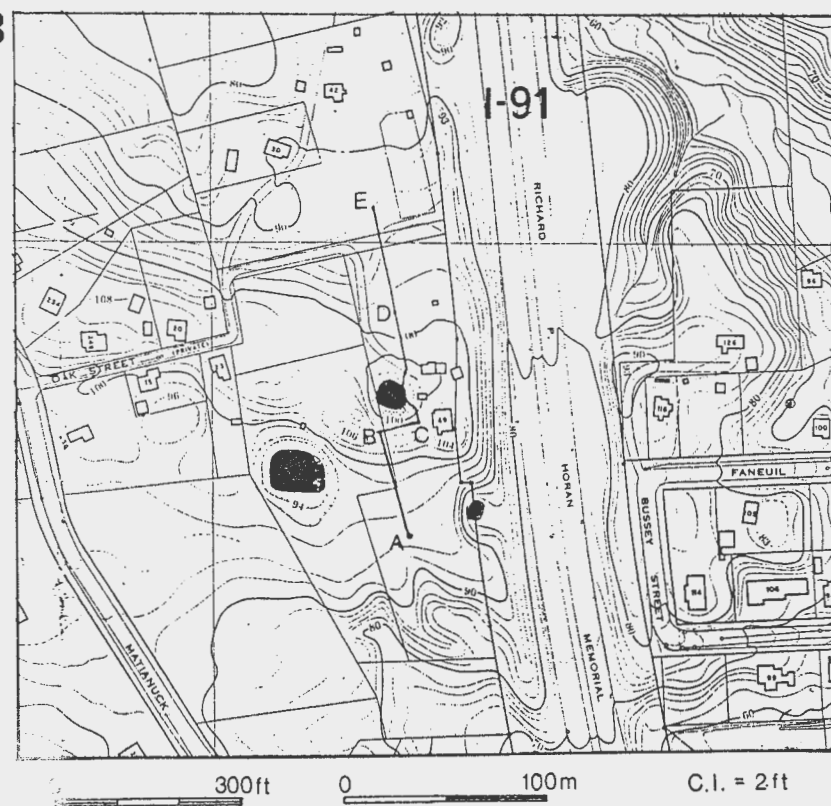


Figure 5. A) Measured Section at I-91 roadcut north of Hartford CT. B) Section of MDC Map 91 (1977 ed.) showing topography of area and location of section line (A - E).

Age and Paleobotanical record

Three ^{14}C dates were obtained from fill material of the I-91 pingo scar. The lower part of the eolian sand body contained detrital wood fragments which yielded a date of **14,330 \pm 430 yr BP** (Beta-35211); the wood was identified as *Salix* (willow). The upper lacustrine material just below the sand fill also contained plant debris. As yet we have not obtained a date on this material, but plant macrofossils identified by Lucinda McWeeney, Yale University include: Twigs- *Vaccinium oxycoccos* (small cranberry), *Vaccinium uliginosum* (bilberry); Seeds- *Empetrum nigrum* L. (crowberry), *Potamogeton* sp (flat stemmed pondweed), *Ranunculus* sp (buttercup), *Vaccinium* sp (mountain cranberry or alpine bilberry), *Cyperus* sp (flat sedge), *Carix* sp (sedge). It is possible that the plant material in the eolian pingo-scar fill may have been redeposited from the upper lake beds during the uplift and collapse of the pingo deformation.

A sample from the lowermost section of the peat body yielded a date of **11,890 \pm 130 yr BP** (Beta-34820). A second (AMS) date of **12,050 \pm 110 yr BP** was obtained from the basal peat sample on several spruce needles and a cone bract. Macrofossil plants and pollen types (identified by L. McWeeney and D. Peteet) include: Needles- *Abies balsamea* (balsam fir), *Picea* sp (spruce), cf *Juniperus* (juniper), *Larix* (larch); Seeds- *Picea* sp (spruce); Bracts- *Abies balsamea* (balsam fir), *Betula* sp (birch); Leaves- *Vaccinium* sp (bilberry or cranberry); Twigs- *Larix* (larch), *Betula* sp (birch), Ericaceae- heath family, *Myrica gale* (sweet gale), *Shepherdia canadensis* (soapberry); Other small plants- *Gaultheria procumbens* (mountain tea), *Vaccinium oxycoccos* (cranberry), *Cyperaceae* (sedge); Other plant parts- Nymphaeaceae (rhizome fragments-water lily), *Equisetum* sp (stems-horsetail), Gramineae- stems grass sp, *Chara fragilis* (green algae), *Mycorrhizal sclerotia* (fungi); Animal- beetle exoskeleton fragments, *Daphnia* (water flea); Pollen- poplar, birch conifer, willow, alder, and spruce.

Four ^{14}C dates were obtained by consultant geologists with EBASCO who were investigating the possibility that the rimmed depressions might be of seismogenic origin. These dates were on samples from vibracores through the fill of pingo scar F, a 50-m-diameter depression located approximately 1 km west of the I-91 cut (STOP 3). Five cores were obtained; GPR lines across the depression aided in interpretation of the vibracores. Core F-7 located in the center of the depression penetrated to a depth of 4 m, and revealed 2.7 m of brown/black peat with minor clastic sediment over 1 m of thinly bedded yellowish brown very fine sand/silt with minor peaty interbeds over 0.3 m of nonbedded gray silt. The lowest thin peat layer in the core yielded a date of **12,630 \pm 240 yr BP** (Beta-46514); the base of the continuous peat body yielded a date of **12,350 \pm 110 yr BP** (Beta-46513). Two other shallower cores F-3 and F-5 penetrated about 2 m into the peat body; a date of **8,325 \pm 100 yr BP** (GX-17050) was obtained from core F-5; a date of **7,850 \pm 220 yr BP** (GX-17053) from core F-3.

A date of **12,200 \pm 250 yr BP** (W-828) was reported by Colton (1960) from "a small peat bog" excavated during construction of the NE runway at Bradley International Airport; this feature is now known to have been one of the many pingo scars in that area.

Genesis of drained Lake Hitchcock Pingo Scars

We suggest the following scenario at the time of pingo formation on drained Lake Hitchcock bottom surfaces probably between 14 and 13 ka. The ice had left the area several thousand years earlier, but the climate remained cold; cold enough to support at least discontinuous permafrost in areas favorable to its development, such as drained lake beds (Mackay, 1979, 1988a, 1988b). Permafrost began to develop on the supersaturated lake-bottom surfaces as soon as the lake drained (fig. 6). As permafrost aggraded downward in the lacustrine sediments pore water was put under hydrostatic pressure. Hydraulic pipes piercing the permafrost released the pressure by localized upward migration of water which turned to ice at the base or within fractures in the permafrost. The repeated addition of ice mass by freezing ground water and the 9% expansion due to the freezing process gradually built a dome of ice causing the sediment cover to be stretched thin over the top. Mass movement in the active layer during summer months moved some of the sediment downslope.

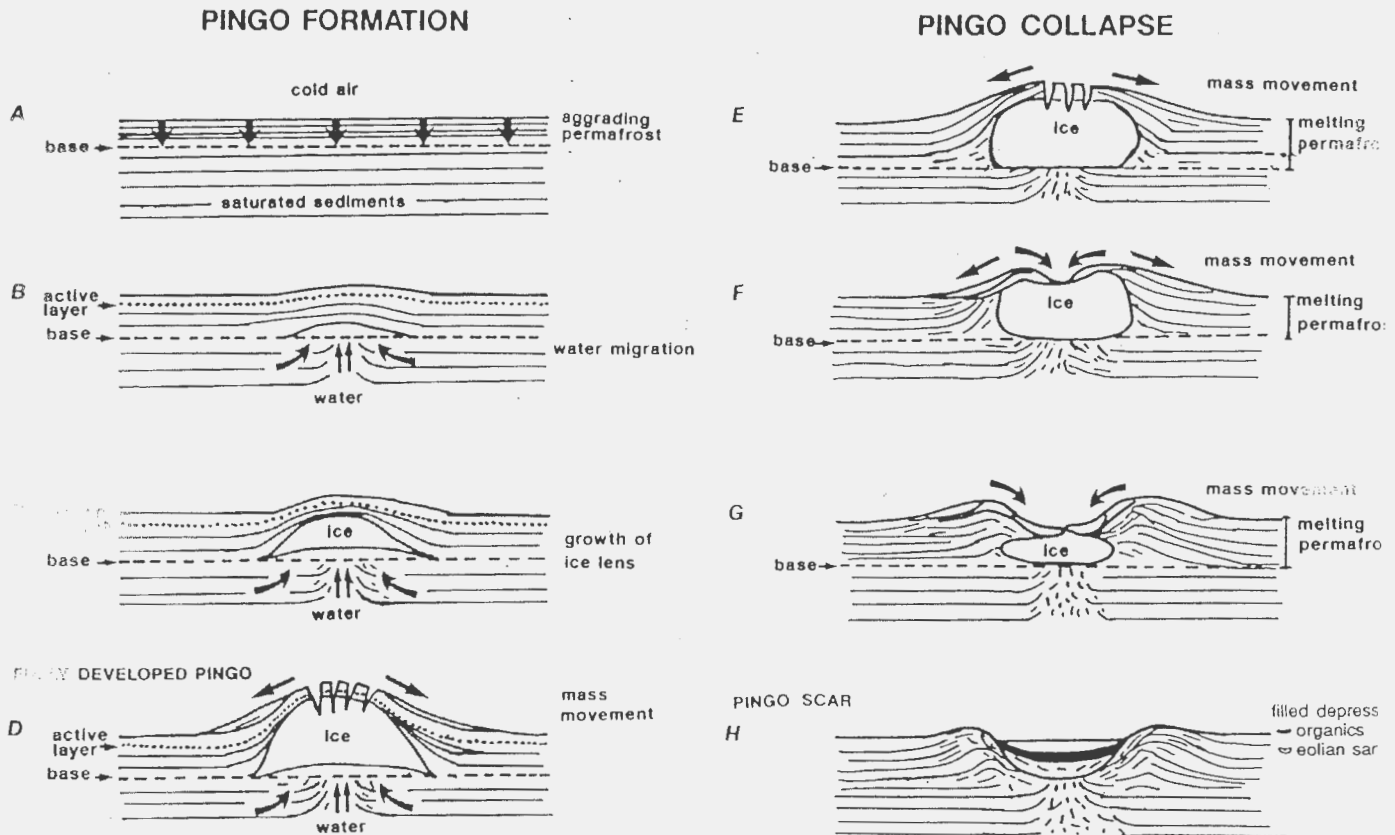


Figure 6. Pingo formation and collapse.

Aggrading permafrost (A) places ground water under pressure (B) which is released through pipes (C) and accumulates (freezes) at the base of permafrost and in cracks within the frozen sediment (D). Sediments immediately below the pingo are deformed and sediments on the surface (in the active layer) are subjected to mass movement. Mass movement continues both into and away from the pingo dome as the permafrost and pingo ice core melt from the surface downward (E, F, and G). The pingo scar (H) is a bowl-shaped depression enclosed by a raised rim which later may be partially filled with eolian sediment or organic material (peat).

Based on size and shape of modern analogs, the 20 to 40-m-diameter pingos might have grown to be 6 to 15 m high. Pingo growth rates on catastrophically drained lakes on Tuktoyaktuk Peninsula, Northwest Territories (Mackay, 1990) yielded a minimum time period of 100 years to generate a pingo of 15 m height.

Before about 12.5 ka the climatic regime had changed such that the pingo ice cores melted creating an inversion of topography (fig. 6E-H). The rampart that built up at the base of the ice-covered dome by colluvial processes became the rim enclosing the depression. Soon after the collapse, wind-transported sand partially filled some of the ponded depressions, followed by predominantly organic deposition after 12.5 to 12.0 ka.

THE DRAINAGE OF GLACIAL LAKE HITCHCOCK

Lake Hitchcock was the longest-lived of the southern New England glacial lakes and the inference of permafrost continuing until after lake drainage has important postglacial paleoclimatic implications. Because the dam for the lake was composed of sediment rather than ice, and its spillway was across a bedrock drainage divide rather than the drift dam (fig. 1), the lake persisted for several thousand years. The life of the lake began with the ice-marginal emplacement of the Cromwell-Rocky Hill delta complex (i.e. drift dam) graded to glacial Lake Middletown, a precursor in the valley to Lake Hitchcock. The details of the early lake history have been presented in Stone and others (1982) and Koteff and others (1988). Correlation of regional ^{14}C dates (Stone and Borns, 1986) places the ice margin in central Connecticut and the inception of Lake Hitchcock at about 16 ka. Sequential ice-marginal deltas graded to Lake Hitchcock record its existence in the valley during the time of ice retreat from central Connecticut to northern Vermont, a distance of about 300 km. Ice-marginal deltas in the southern part of the lake record high lake levels, between 115 ft (35 m) and 90 ft (27 m) in altitude, at the spillway; meteoric-water-fed deltas in the south and ice-marginal deltas from Chicopee, MA northward record a stable lake level at 82-ft (25 m) in altitude at the spillway (Koteff and others, 1988). The lake drained due to breaching of the drift dam at a time after the ice margin had left the northerly reaches of the Connecticut valley.

For many years, the time of Lake Hitchcock drainage was accepted to be approximately 10.7 ka based on dates and interpretations presented by Flint (1956). The five ^{14}C dates >12.0-14.0 ka from basal organic fill of the rimmed depressions on the Hitchcock lake bottom clearly demonstrate lake drainage before that time (12.6 ka); the depressions (regardless of their mode of origin) could not have formed until after the lake drained. The ~14-ka date on detrital wood in the eolian pingo-scar fill (I-91 cut, fig. 5) possibly indicates an even earlier time of drainage, but the twigs may have been redeposited in the fill from underlying lake beds. A recent exposure of upper lake beds 1 km north of the I-91 cut in Windsor, CT (Matianuck Ave. site, fig. 10) has provided more evidence for a ~14-ka time of lake drainage.

Matianuck Avenue Site, Windsor, CT

Although the lake-bottom surface in the area of the Matianuck Ave. site displays many pingo-scar depressions, the section of the roadcut available at the time it was studied did not cross any of these features. It did reveal approximately 4 m of upper lake-bottom section (see fig. 10, Stop 4). Thin layers of plant debris interbedded with fine sand and silt layers were found 0.5 m above varved beds and 2.5 m below the top of the section. Small plant macrofossils in this material were exceptionally well preserved (see below). Two ^{14}C dates were obtained on the plant material, **14,120±90 yr BP** (Beta-52711) and **13,080±100 yr BP** (W-6397); both dates came from the same sample of material, so the reason for the 1,000-yr difference between the two dates is not clear at present; a willow twig submitted by Norton Miller (see below) for AMS dating should resolve the problem. Current directions of crossbeds and ripples and composition indicate that these beds are distal from a mapped post-stable-stage delta about 2 km to the northeast, built into Lake Hitchcock by the paleo Farmington River (Stone and others, 1992; Stone and Schafer, in prep.). This delta records lowering of the lake by about 5 m from the stable 82-ft (25-m) level; lowering of this amount produced a lake shoreline within a few hundred meters of the Matianuck site. It is probable that the post-stable delta represents lake-level lowering due to initial breaching of the Cromwell-Rocky Hill drift dam.

Article D. GLACIAL LAKE HITCHCOCK

excerpted from the Quaternary Geologic Map of Connecticut and Long Island Sound Basin (Stone and others, 1998; U.S. Geological Survey Open File Report OF 98-371)

Glacial Lake Hitchcock existed in the upper Connecticut River basin in Connecticut, Massachusetts, Vermont, and New Hampshire, lengthening to at least 185 mi as the ice retreated northward to the vicinity of Burke, Vt. The Connecticut River valley was dammed to an altitude of 150 to 160 ft in the vicinity of Rocky Hill and Glastonbury by deposits of glacial Lake Middletown (**Mc** and **db**); this mass of stratified drift is often referred to as "the Rocky Hill dam." The spillway for Lake Hitchcock was not over the dam, however, but at the lowest place across the Mattabesset River drainage divide between the Hartford basin and the Middletown-Berlin basin in New Britain. When the ice margin first retreated into the Hartford basin, north of that divide, glacial Lake Middletown water covered the later New Britain spillway location and early ice-marginal deltas in the Hartford basin were controlled by glacial Lake Middletown. Not until glacial Lake Middletown had dropped to below 115 ft could the New Britain spillway area emerge and glacial Lake Hitchcock exist as a separate water body; this occurred at about the time that the ice margin was at Windsor and East Windsor.

During the early life of glacial Lake Hitchcock, the New Britain spillway was eroded into till and older stratified drift so that water levels at the spillway dropped from about 115 ft to 82 ft in altitude (Langer, 1977; Langer and London, 1979). In Connecticut, all ice-marginal and distal meltwater-fed deltas, as well as one small delta built by meteoric water, record lake levels higher than the longer lived stable level. These deltas show a gradual lowering of lake level as the ice retreated northward and the New Britain spillway was incised down to bedrock. Ice-marginal deltas in Windsor (**Hhw**) and East Windsor (**Hhe**) record 110- to 115-ft levels at the spillway. To the north, ice-marginal deltas in Suffield (**Hhr**) and Enfield (**Hhs**) indicate 105- to 110-ft levels at the spillway; still farther north in Suffield and Enfield, Shea Corner (**Hhc**) and Enfield (**Hhn**) deltaic deposits of record levels just below 100 ft at the New Britain spillway. This early phase of glacial Lake Hitchcock is recorded by ice-marginal deltas that are found well into southern Massachusetts and that were built to lake levels between 85 and 95 ft at the spillway. This higher-than-stable-level phase of the lake is referred to as the "Connecticut Phase" (Koteff and others, 1988). It is important to note that deepening of the spillway channel was controlled by conditions 30 to 40 mi to the south, because earlier workers have always considered that the New Britain spillway was an independent control for lake levels. Base level for waters exiting the spillway was controlled by down-cutting in the lower Connecticut River valley and by lowering levels of glacial Lake Connecticut in

the Long Island Sound Basin. The Rocky Hill dam area was glacio-isostatically depressed about 145 ft and the New Britain spillway area was depressed about 165 ft more than the area at the mouth of the Connecticut River. In order for the New Britain spillway to lower by 33 ft during the early phase of the lake, glacial Lake Connecticut had to have already lowered to below -82 ft (-25 m) in altitude.

Delta levels in Massachusetts indicate that a stable lake level, 82 ft in altitude, had been reached by the time the ice margin had retreated to just north of the Chicopee River valley; regional correlation of ^{14}C dates (Stone and Borns, 1986) place the ice front in this position at about 15 ka. The 82-ft level indicates that the water flowing through the spillway was about 24 ft deep because its bedrock floor today is at about 58 ft in altitude. Altitudes of topset-foreset contacts of ice-marginal deltas, from southern Massachusetts to the lake's northernmost extent, project to the stable level (82 ft at the New Britain spillway) on a straight line which is tilted up to the north-northwest at a slope of 4.74 ft/mi. The linearity of these projected delta altitudes indicates that the lake level was stable during the time of ice retreat from Chicopee, Mass., to Lyme, N.H., and that postglacial rebound of the land surface did not begin until after all ice-marginal deltas had been built, probably between 14 and 13.5 ka (Koteff and Larsen, 1989). Deltas that were not associated with the ice margin, but rather were built by meteoric water in most river valleys that entered the lake, also project to the stable lake level. In Connecticut, these include unit **H1h** associated with the Hockanum River, unit **H1s** associated with the Scantic River, and unit **H1b**, where the Farmington River constructed a large delta northeastward into the lake in the area now surrounding Bradley International Airport. The Bradley International Airport delta covers about 20 mi² and the fact that its entire surface (which is tilted up to the N. 21° W. in the amount of 4.74 ft/mi) is graded to the stable 82-ft level provides evidence for the long duration of the stable level and also indicates that the lake was not affected by glacio-isostatic tilting until after nearly all of its deltas had been constructed.

It is important also to note that the New Britain spillway could not have lowered further than the 82-ft level. This is because the 82-ft altitude at the New Britain spillway is equivalent to a -82-ft altitude at the mouth of the Connecticut River when the 164 ft of differential depression between the two localities is taken into account. The base of the channel, through which the paleo-Connecticut River carried water that spilled from glacial Lake Hitchcock, was imposed on bedrock at -89 ft (-27 m) in altitude at the mouth of the present Connecticut River east of Saybrook Point; this point was the actual control for the "Stable Phase" (Koteff and others, 1988) of glacial Lake Hitchcock. The "Stable Phase" of glacial Lake Hitchcock lasted from about 15 ka until about 13.7 ka; during this time, the southern part of the basin (south of the Holyoke Range in Massachusetts) was largely filled with deltaic and lake-bottom sediments. Preserved lake-bottom surfaces in

Connecticut are at about 45 ft in altitude in the south and 145 ft in the north; the tilted stable-level paleowaterplane over this area is at 63 ft in altitude at the north edge of the Rocky Hill dam and 172 ft at the Massachusetts border; thus, toward the end of the "Stable Phase" before the dam was breached, water depths in the lake were only 20 to 25 ft. Because the bedrock basin that contained the lake north of the Holyoke Range in Massachusetts is deeper, the lake was not filled with sediment to the extent that it was in the southern basin. North of the Holyoke Range in Massachusetts, preserved lake-bottom surfaces are at 150 ft in altitude, and at the end of the "Stable Phase" water depth was about 150 ft.

Fluviodeltaic deposits (**ft** and **hf**) built southeastward into the lake by the Farmington River record a "Post- stable Phase" (Koteff and others, 1988) of the lake during which levels were lower than the 82-ft level at the New Britain spillway. A topset-foreset contact in the **Hf** deltaic deposits north of the Farmington River is at 127 ft; delta-surface altitudes in the same unit to the south of the river indicate slightly lower water levels. These levels project southward below the New Britain spillway level to 50 to 60 ft in altitude at the Rocky Hill dam and record lowering of lake levels as the dam was entrenched. A preserved 55-ft terrace inset into the Rocky Hill dam sediments on both sides of the present Connecticut River in Rocky Hill and Glastonbury records this "Post-Stable Phase" which was relatively brief in Connecticut. A ^{14}C date $13,540 \pm 90$ B.P. (Beta-59094, CAMS-4875) on plant debris in lacustrine sands at the top of the lake-bottom section (radiocarbon-dated locality #9) associated with the Farmington River deltaic deposits (**Hf**) delta establish that the time of dam breach was at about 13.5 ka.

The dam most likely was breached by headward erosion of streams on its south side, possibly by ground-water sapping and possibly aided by earthquakes generated by the initiation of postglacial rebound. Regardless of the mechanism by which the dam was breached, glacial Lake Hitchcock could not lower below stable level, much less drain, until its bed was raised by glacio-isostatic tilting. Dam breaching and initiation of isostatic rebound was required in order to establish the lower water-level altitudes recorded in the "Post-Stable Phase" Farmington River deltaic deposits (**Hf**). Once this process began, it proceeded rapidly as the dam was incised from just above 60 ft in altitude (the stable level at the dam) to just above 40 ft; once this 20 ft of lowering was accomplished, glacial Lake Hitchcock, south of the Holyoke Range, was entirely drained and the newly formed Connecticut River began to incise the lake floor (along the terraces of unit **st**) over the 50-mi stretch between the Holyoke Range and the breached dam. Glacial Lake Hitchcock continued to exist north of the Holyoke Range with initial water depths of about 130 ft (lowered from stable level by only 20 ft); continued lowering of the lake was controlled by the rate of rebound, which made it possible for the lake bed south of the Holyoke Range to be incised.

An approximate 4,000-year life span for glacial Lake Hitchcock was indicated by Antevs (1922) through a method of correlating varves in clay pits from Hartford, Conn., to the north end of the lake basin in St. Johnsbury, Vt. This method assumes that the silt-clay varve couplets are annual summer and winter layers and that regional seasonal fluctuations affected the thickness of individual varves over the entire lake basin. Varved silts and clays of glacial Lake Hitchcock were used to construct Antevs' (1922) New England varve chronology between varve-year 3,001 and varve-year 7,000. Recently, Ridge and Larsen (1990) fit a 533-year varve section from Canoe Brook in southern Vermont into the relative varve chronology of Antevs (1922); they also placed the chronology in an absolute time frame with a 12.4 ka ^{14}C date on plant debris in the Canoe Brook section at the position of varve 463 (varve 6,150 in the Antevs chronology). Using this calibration of the varve chronology, lacustrine deposition at the south end of glacial Lake Hitchcock (varve 3,001) began at about 15.5 ka. The early Connecticut phase was followed by the longer stable phase of the lake which lasted until about 13.5 ka (varve 5,050). The post-stable phase of the lake, which lasted only briefly in Connecticut, continued for another 2,000 years north of the Holyoke Range until about 11.5 ka (varve 7,000) (Stone and Ashley, 1992, 1995; Stone, 1999).

POSTGLACIAL CONDITIONS

Postglacial deposits in Connecticut include stream-terrace (**st**), talus (**ta**), dune (**d**), floodplain alluvium (**a**), swamp (**sw**), salt-marsh (**sm**), beach (**b**), fluvial-estuarine channel-fill (**ch**), and marine delta (**md**) deposits; the onset of postglacial conditions was time-transgressive and began several thousand years earlier in the southern part of the State than in the northern parts.

In the Long Island Sound Basin, significant postglacial events include the drainage of the glacial Lake Connecticut and subsequent sea-level rise. The remnant glacial lake was probably completely drained by 15.5 ka and a fluvial channel system (linear scarp symbol on map) was being carved on the lake floor by meteoric streams flowing to the south in coastal Connecticut, and to the north on the north shore of Long Island; these tributary channels joined a major east-west trending trunk channel which also received distal meltwater drainage from the Hudson River valley to the west (Stanford and Harper, 1991). The channel system exited the Basin through the lake- spillway notch in the end moraine at The Race and provided a path through the moraine for transgression of the sea from the south (Lewis and Stone, 1991). Minor fluvial sediments were deposited in the bottoms of the channels during the time that they were occupied by streams. The channels are filled predominantly with estuarine sediment (**ch**) deposited as the early postglacial sea flooded these low-lying areas of the drained lake basin when eustatic sea-

level began to rise significantly between 15 and 16 ka (Fairbanks, 1989; Bard and others, 1990) and before glacio- isostatic rebound began.

A major wave-cut marine unconformity (**mu** on seismic section **C-C'**) was cut across the top of the estuarine channel fill and over higher lake deposits as sea level rose. The marine unconformity is present in seismic sections up to altitudes of about -25 m, indicating that sea level probably rose to this height in central Long Island Sound before crustal rebound began.

Figure 7 shows a conceptual relative sea-level curve for central Long Island Sound (highlighted line). The curve was derived by combining the glacio-eustatic sea-level curve from Barbados (Fairbanks, 1989; Bard and others, 1990) with a curve representing the timing and total depth of glacio-isostatic depression in central Long Island Sound. The uplift curve is based on several assumptions (listed on diagram) that are indicated from regional evidence, some of which is presented in this report and in others (Koteff and Larsen, 1989; Stone and Ashley, 1995). The presence of the extensive marine delta (**md**) that records a -40-m relative sea level in central Long Island Sound (Lewis and Stone, 1991; Stone and Lewis, 1991) provides good evidence for the conceptualized relative sea-level curve. The large volume of delta sediment required a significant length of time for construction and the constant -40 m depth of the topset-foreset contact indicates that relative sea level was stable during the deposition of the delta. The only possible source of the great volume of sediment contained within the marine delta was the drained lakebed of glacial Lake Hitchcock in the Connecticut valley to the north. This sediment supply became available only when the stable phase of Lake Hitchcock ended at about 13.5 ka; as previously discussed, glacio-isostatic uplift had to occur in order for glacial Lake Hitchcock to drain. Regional evidence from northern New England (Barnhardt and others, 1995; Koteff and others, 1993; Koteff and others, 1995) also indicates that isostatic rebound began around this time. Thus, the early rapid rate of uplift was balanced with the equally rapid rate of eustatic sea-level rise resulting in a sea-level stand in Long Island Sound at about -40 m for several thousand years (between 13.0 and 9.5 ka). During that time, the marine delta was built and the Connecticut River terrace and floodplain surfaces were incised. Recently obtained ^{14}C dates (9,370 \pm 100 Beta-52257, 8,530 \pm 80 Beta-52256) on basal organic material beneath the lowest terrace surfaces along the Connecticut River in Massachusetts indicate that most of the postlake

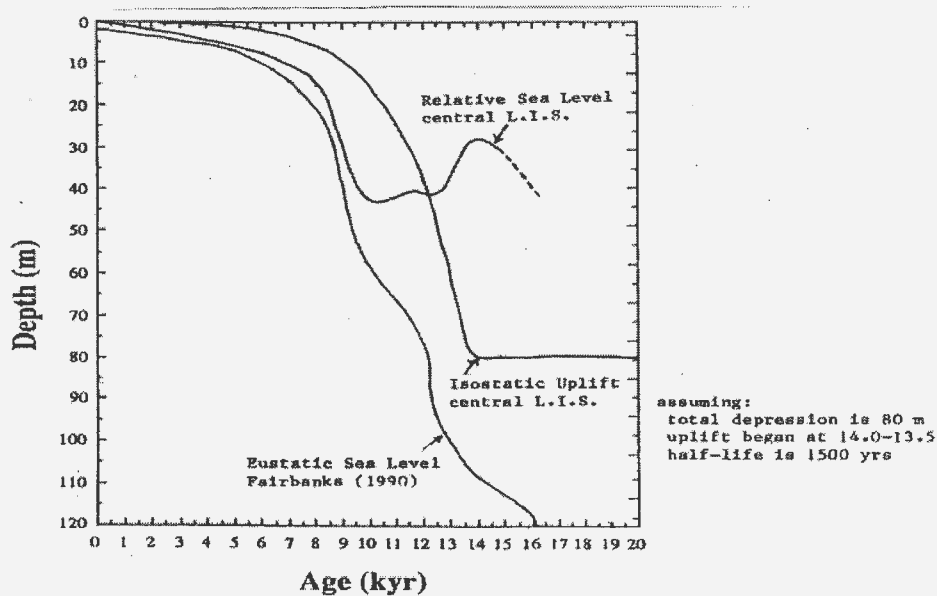


Figure 7. Conceptual relative sea level curve for central Long Island Sound.

incision into the lakebed had been accomplished by ~9.0 ka (Stone and Ashley, 1992). The volume of eroded lakebed sediment, as calculated from the area and depth of incised terraces, is 12 billion m^3 ; this material now composes the marine delta, the calculated volume of which is 11.5 billion m^3 .

As eustatic rise overtook the rate of isostatic rebound, relative sea level in central Long Island Sound rose continuously; the transgression submerged the marine delta and a blanket of marine mud (seismic unit **mm** shown only in section **C-C'**) accumulated over the entire basin. As marine waters deepened, intense tidal-scour conditions developed in eastern Long Island Sound, resulting in the local reworking of marine-delta sediments and the development of a very large sand-wave field in that part of Long Island Sound (Fenster and others, 1990). A record of 4 to 5 m of sea-level rise during the last 4,000 to 5,000 years is preserved in coastal salt-marsh deposits (Bloom and Stuiver, 1963; van de Plassche and others, 1989; Patton and Horne, 1991; van de Plassche, 1991).

In most of mainland Connecticut, postglacial activity consisted predominantly of incision of glacial deposits by meteoric streams along stream-terrace surfaces, followed by the establishment of floodplains at modern levels. Streams had eroded to modern floodplain levels relatively early, in some cases before 12.0 ka (O'Leary, 1975; Stone and Randall, 1978). Postglacial winds were intense and widespread as indicated by the ubiquitous blanket of eolian sand and silt that overlies glacial sediments throughout the State and in which the modern soil is developed. The postglacial climate was severely cold for several thousand years following deglaciation. Paleobotanical studies reveal that treeless, tundra vegetation dominated by dwarf willow (*Salix herbacea*), sedges (*Cyperus* and *Carix*), herbs and shrubs (*Dryas*, *Artemesia*), dated from earlier than 15 ka to about 13 ka, was present in the area (Davis and others, 1980; Gaudreau and Webb, 1985; Jacobson and others, 1987; Thorson and Webb, 1991). Also, wedge-shaped features with a polygonal-ground pattern, interpreted as ice-wedge casts, deform eolian-sand-capped glacial sediments in numerous localities in Connecticut (Schafer and Hartshorn, 1965; Schafer, 1968; O'Leary, 1975; Stone and Ashley, 1992,). These features indicate that permafrost existed locally in areas where substrate conditions were favorable to its formation. The presence of permafrost structures indicates that mean annual temperatures were below 0°C during the early postglacial time interval.

In the upper Connecticut basin, postglacial conditions were dominated by the continued existence of glacial Lake Hitchcock several thousand years after the ice margin retreated from the area. Extensive fields of eolian sand dunes formed in the treeless environment, indicating the continued effects of strong winds. Dunes are present on the relict deltaic and lake-bottom surfaces of glacial Lake Hitchcock. Dunes on deltaic and high-level lake-bottom surfaces were formed by north to north-northeasterly paleowinds; these surfaces were available as early as 15.5 ka. Dunes on stable-level lake-bottom surfaces were formed by northwesterly paleowinds; these surfaces became available at about 13.5 ka as glacial Lake Hitchcock drained. Evidence that severely cold temperatures persisted until the time of glacial Lake Hitchcock drainage exists due to the presence of hundreds of circular to subcircular, rimmed depressions (interpreted as pingo scars) developed in the drained lakebed sediments (Stone and Ashley, 1989; Stone and others, 1991; Stone and Ashley, 1992). Paleobotanical records indicate a warming of the postglacial climate at about 12.5 ka, accompanied by reforestation of the landscape by successive spruce, pine, and hardwood forests from 12.5 to 9 ka (Davis, 1980; Gaudreau and Webb, 1985; Jacobson and others, 1987).

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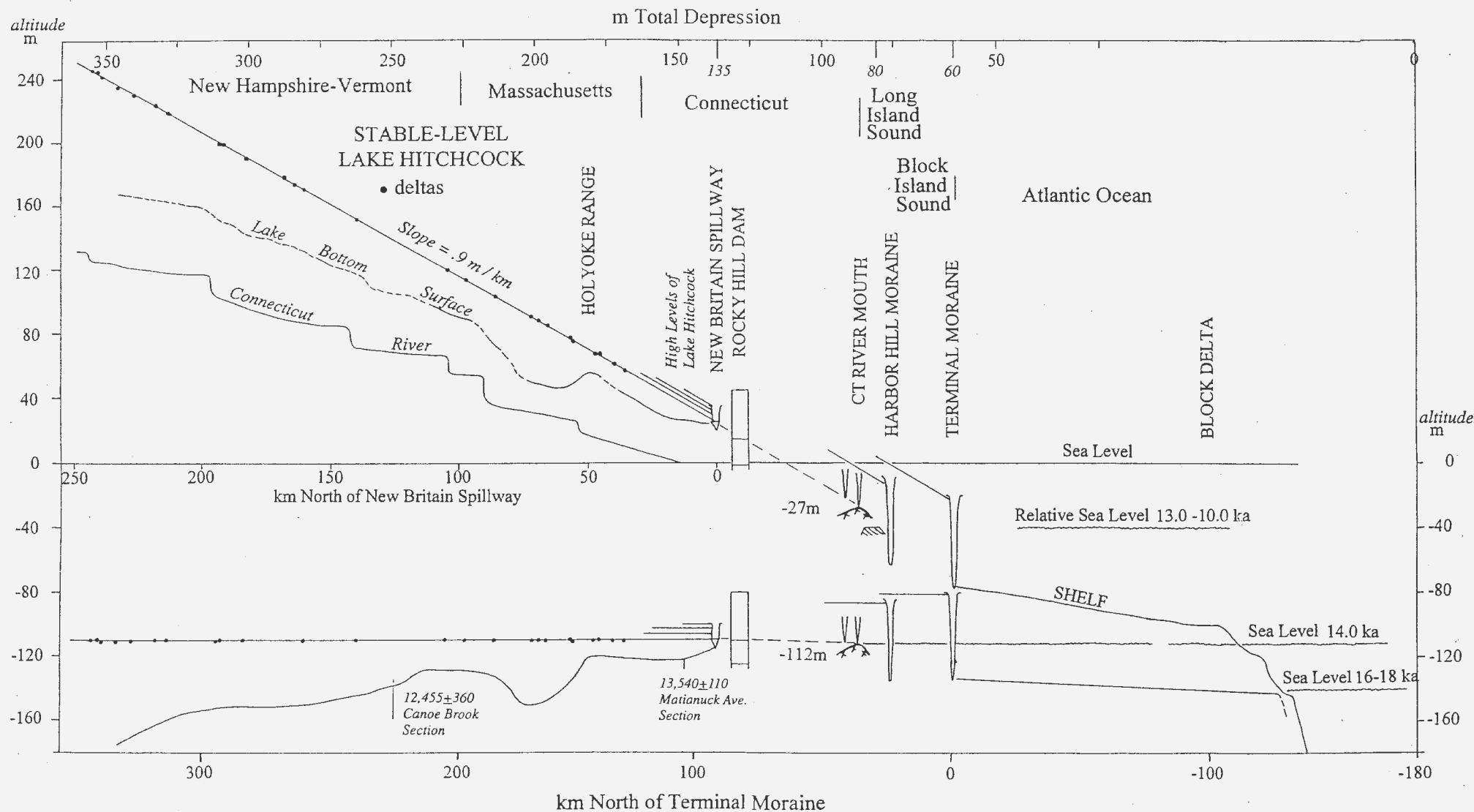


Figure A. Graphical depiction of positions of Lake Hitchcock surfaces and important features south of Rocky Hill Dam. Upper part of diagram shows positions as they occur today; lower part of diagram shows positions after differential depression is reconstructed across the region.

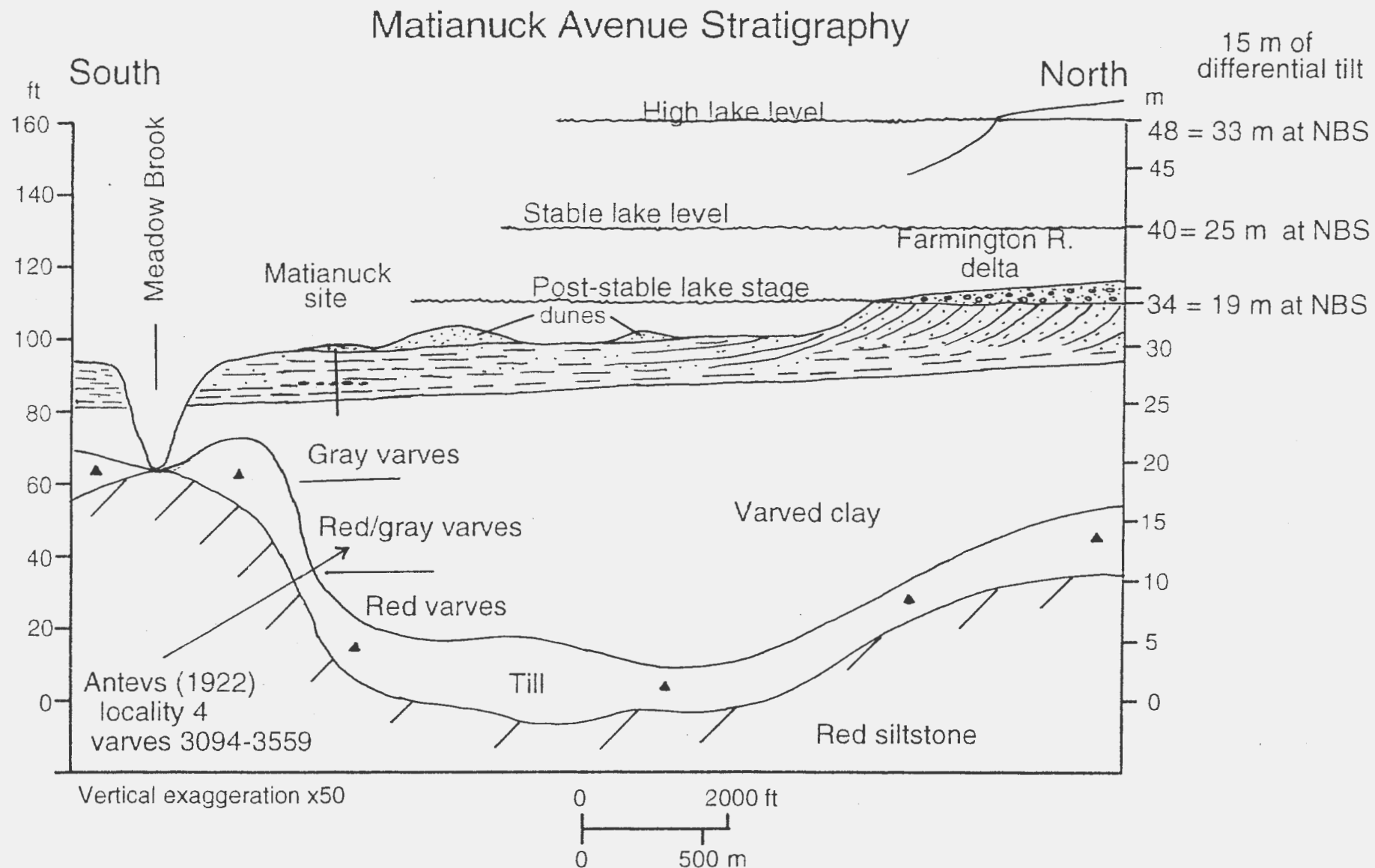


Figure B. Regional cross-section showing stratigraphy at the Matianuck Avenue site in Windsor, Conn., including subsurface extent of gray and red varves as indicated by local test borings, relative position of nearby Anteys (1922) varve locality 4, and lateral continuity with the post-stable stage Farmington River delta.

Article E. ABSTRACTS WITH FIGURES

Stone, J.R. and Ashley, G.M., 1995, TIMING AND MECHANISMS OF GLACIAL LAKE HITCHCOCK DRAINAGE: Geological Society of America Abstracts with Programs, v. 27, no. 1, p. 85.

Glacial Lake Hitchcock (GLH) and its predecessors, Lake Connecticut and Lake Middletown dominated the deglacial history of Long Island Sound Basin and the Connecticut River Valley for thousands of years following ice retreat from the terminal moraines on Long Island. GLH was the longest-lived of these lakes, recorded by a 300-km-long complex of ice-marginal deltas, meteoric-stream-fed deltas, and varved lake-bottom sediments that extends from the drift dam at Rocky Hill and the spillway at New Britain, Conn. to the vicinity of Lyme, N.H. During early high levels of the lake, the spillway channel was incised about 10 m through drift to bedrock. During the long-lived stable stage of the lake, water level at the spillway was at 25 m altitude. Subsequent post-stable stages of the lake existed at lower levels. The post-stable stage lasted only briefly south of the Holyoke Range in Mass., where the stable stage was relatively shallow, and was accompanied by breach of the drift dam and abandonment of the New Britain spillway. A longer post-stable stage existed for about 2,000 years in the northern part of the basin, where the lake was deeper. Water that spilled from the northern basin carved a broad, 75-km-long channel in the drained lake bed to the south and exited the basin through the breached dam. Sediment eroded from the incised lake bed was deposited as an extensive marine delta in central Long Island Sound at a time when relative sea level stood at -40 m. Detailed mapping from marine seismic-reflection profiles in Long Island Sound, Block Island Sound, and the Connecticut River estuary coupled with regional synthesis onland have revealed features significant to the history of GLH, including the buried paleochannel of GLH drainage through the lower Connecticut River valley, across Long Island and Block Island Sounds, and through the terminal moraine. With consideration of glacio-isostatic depression over the entire drainageway south of GLH (*see inserted figure A*), it is evident that this lake never existed independently; levels were always determined by base levels to the south. During the stable stage of GLH, before rebound took place, the stream profile south of the spillway had an extremely low gradient (<0.00005) and was directly connected to sea level in the Connecticut River estuary that stood at a relative altitude only 2 m lower than the lake level to the north. Under these conditions, it was impossible for GLH to lower below the stable level until glacio-isostatic rebound began. The time at which GLH lowered 4-5 m below stable level (and hence the time of initiation of glacio-isostatic rebound in southern New England) is recorded by a new AMS ^{14}C date (13,540 \pm 90 BP; Beta-59094/CAMS-4875) obtained on plant fragments in lacustrine bottomset beds of the GLH Farmington

River post-stable delta at the Matianuck Ave. site in Windsor, Conn. (*see inserted figure B*), described by Stone and Ashley (1992).

Stone, Janet R., 1999, EFFECTS OF GLACIO-ISOSTACY AND RELATIVE SEA LEVEL ON LATE-GLACIAL AND POSTGLACIAL WATER LEVELS IN THE CONNECTICUT RIVER VALLEY AND LONG ISLAND SOUND: Geological Society of America, Abstracts with Programs, v. 31, no. 2, p. 71.

Glacio-isostatic depression and the position of sea level off the continental shelf south of Long Island and in Long Island Sound (LIS) during early retreat of the late-Wisconsinan ice sheet had a profound effect on the water levels of Glacial Lake Hitchcock (GLH). The lake occupied the Connecticut River valley between about 15.5 and 11.5 ka (radiocarbon years), during which time, lake-level altitudes fell from early high levels between 35 and 26 m, through a long stable period at 25 m, to post-stable levels below 25 m after the Rocky Hill dam was breached. Graphical analysis (*see inserted figure A*) of these ancient water levels is based on present altitudes of the New Britain spillway, stable-level deltas, and lake-bottom surfaces, and base-level controls at the mouth of the Connecticut River and notches through the terminal moraines. All altitudes were adjusted for the total (absolute) amount of glacio-isostatic depression, which is defined by a plane sloping 0.9 m/km to the NNW. The total amount of depression is calibrated from the ~80 m required for the construction of a marine delta at -40-m altitude in central LIS as GLH drained. When surfaces across the entire drainage system are restored to the pre-uplift configuration, it is clear that levels of GLH were ultimately controlled by relative sea level far to the south of the New Britain spillway, and that the lake could not drain until postglacial uplift began. The Rocky Hill dam was breached by 13.5 ka, as recorded by a ^{14}C date from bottomset beds of a post-stable delta just north of Hartford, Conn. (*see inserted figure B*). Restoration of the pre-uplift lake-bottom surface reveals that initial lowering of lake level by only 10-12 m resulted in the complete disappearance of the lake south of the Holyoke Range in Massachusetts, while a deep lake continued to persist in the northern basin until at least 12.4 ka (as recorded by a ^{14}C date at Canoe Brook, Vt). GLH in the northern basin drained across the exposed lake-bottom surface to the south; this surface was entrenched and terraced as uplift progressed, and eroded sediment was deposited as a marine delta in a -40-m relative sea level stand in central LIS.

Most of the total amount of glacio-isostatic depression was recovered by about 9.0 ka, and the ancestral Connecticut River had entrenched into lake sediments down to levels near the modern floodplain by 8.5 ka. A postglacial sea-level curve for central LIS shows that the rate of glacio-eustatic sea-level rise equaled the rate of crustal uplift, resulting in a stable relative sea level stand at -40 m between 13.5 and 9.5 ka. By about 9.0 ka, the rate of eustatic rise began to exceed the rate of uplift, and sea level in LIS rose rapidly from the -40-m stand.

El Niño–Like Climate Teleconnections in New England During the Late Pleistocene

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A glacial varve chronology from New England spanning the 4000-year period from 17,500 to 13,500 calendar years before the present was analyzed for evidence of climate variability during the late Pleistocene. The chronology shows a distinct interannual (3 to 5 years) band of enhanced variability suggestive of El Niño–Southern Oscillation (ENSO) teleconnections into North America during the late Pleistocene, when the Laurentide ice sheet was near its maximum extent and climatic boundary conditions were different than those of today. This interannual variability largely disappears by the young end of the 4000-year chronology, with only the highest frequency components (roughly 3-year period) persisting. This record provides evidence of ENSO-like climate variability during near-peak glacial conditions.

The unusual recent behavior of ENSO, including the exceptionally strong warm events of 1982–83 and 1997–98 and the predominance of El Niño–like conditions during the past two decades (1), has highlighted the possibility that ENSO may be influenced by anthropogenic forcing (2). The limited success of models in predicting El Niño events (3) and the nonstationary statistical character of ENSO (4) and its teleconnections (5, 6) over the past two centuries indicate that our understanding of ENSO from the short instrumental record is incomplete. Thus, there is considerable interest in the long-term evolution of ENSO and its relation to low-frequency climate forcing.

Paleoclimate studies using tree rings, corals, and ice cores support the existence of century-scale irregularities in the character of ENSO variability (6, 7). For example, during the late 17th through mid-19th century, decadal ENSO-like variability persisted and interannual ENSO variability was strongly muted (6–8). The fact that this interval corresponds to a period of substantially lowered solar irradiance (the “Spörer Minimum”) has prompted speculation that solar forcing may influence the amplitude of interannual ENSO variability (6). Similarly, models suggest that greenhouse radiative forcing may lead to decreased amplitude of interannual ENSO variability (2).

Paleoclimate records show fluctuations in the intensity of ENSO variability during the Holocene. Coral isotopic data from the Great

Barrier Reef (9), a region currently influenced by ENSO, indicate warmer conditions with an absence of interannual variability during the middle Holocene, suggesting that ENSO (or its teleconnections to this region) was suppressed or absent at this time. Similar evidence for reduced ENSO variability during the middle Holocene was found in records from Australasia (10). Faunal records off the coast of Peru (11) have been used to suggest that ENSO events are a middle to late Holocene phenomena, but this interpretation is controversial (12). However, storm-induced debris flows at the low-frequency limit of the conventional ENSO band from a lake in Ecuador defy this interpretation and may extend the record of ENSO activity to 7000 calendar years before the present (cal yr B.P.) and possibly to 15,000 cal yr B.P. (13). In addition, spectral analysis of 264 glacial varves from the Lake Huron basin indicates the presence of ENSO-like interannual variability immediately after the Younger Dryas (~12,000 cal yr B.P.) (14). However, additional late Pleistocene records are needed to confirm the presence and nature of ENSO variability before the middle Holocene.

From a dynamical perspective, it is likely that astronomical forcing has modulated the amplitude and frequency of El Niño variability (15). Modeling results (16) suggest that precessional forcing should have sharply reduced the frequency of strong El Niño events before the middle Holocene, although interannual ENSO variability may not have ceased altogether. A recent modeling experiment forced by early Holocene insolation (15) shows a decreased amplitude of low-frequency variability but a persistence of high-frequency ENSO variability in relation to modern climate. It has not yet been possible to examine interannual ENSO-related variability over millennial-scale changes in

insolation and during late Pleistocene glacial conditions.

We analyzed the recently revised New England (NE) varve chronology based on varves from proglacial lakes formed during the recession of the Laurentide ice sheet 17,500 to 13,500 cal yr B.P. (Fig. 1). Thick clastic varves formed because of seasonal changes in sediment deposition (melt season versus lake ice cover), with the coarser melt-season layer contributing most of the variation in varve couplet thickness (17). Annual varve thickness was a function of glacial melting and surface runoff controlled by regional climate, allowing varves to be correlated within and between contemporaneous proglacial lakes throughout New England (18, 19). Ice-proximal and thin ice-distal varves do not accurately represent annual meltwater runoff and regional climate and thus were removed from the chronology. The sensitivity of varve couplet thickness to annual climate in New England, a region known to currently exhibit seasonal ENSO influence (20), makes the NE varve chronology a potentially ideal record for studying changes in ENSO or its teleconnections during the late Pleistocene. Perhaps more relevant than modern ENSO teleconnections in New England is evidence that the extratropical teleconnections of ENSO-related sea surface temperature patterns into this region were likely considerably stronger during glacial times (21). It is thus quite plausible that an analysis of the interannual temperature-related variability evident in this varve chronology might yield information regarding changes in ENSO or its teleconnections during the late Pleistocene.

In 1922, Antevs (18, 19) correlated >4000 varves within Glacial Lake Hitchcock and between contemporaneous glacial lakes in New England to create the original NE varve chronology. Antevs assembled this chronology by creating “normal” curves of average varve couplet thickness measurements from >100 correlated varve exposures. Multiple overlapping sections were measured to eliminate counting and measurement errors. The original NE varve chronology had two gaps within the series, which have now been corrected through paleomagnetic investigations in Vermont (22) and correlation from a core collected in Amherst, Massachusetts (23), yielding a continuous varve chronology (24).

The integrity of the NE varve chronology has been established through a variety of methods. First, many of the varve sections used by Antevs (18, 19) have been remeasured (22, 23); second, varve couplet thickness was measured in sections not used to formulate the original chronology (25); and third, paleomagnetic secular variation investigations (22, 25, 26) were performed to con-

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firm the correlation between overlapping segments. Most important, the previously floating chronology has now been fixed through careful accelerator mass spectrometry dating (22, 23, 25), confirming correlations within the chronology and providing well-constrained numerical dating of the NE varve chronology.

A number of regional varve thickness curves from the Glacial Lake Hitchcock portion of the NE varve chronology, deposited between 17,500 and 13,500 cal yr B.P., were combined to yield the continuous chronology analyzed here, NE varve numbers 2868 through 6900 [after correction for gaps in original chronology, see (22, 23)] (Fig. 2A). Before analysis, the raw varve chronologies were combined to yield a composite series representative of the underlying climate variability. Observations from modern lakes containing varved sediment indicate a logarithmic relation between varve thickness and early summer temperature, which influences meltwater and suspended sediment discharge (27, 28, and references therein). These underlying temperature variations, which typically exhibit a Gaussian distribution, are what we sought to isolate from the varve chronology. Consequently, a lognormal transformation of the varve thickness data was performed to yield a Gaussian-distributed time series appropriate for statistical analysis (Fig. 2B).

To facilitate joining the data sets into one continuous sequence, we removed low-frequency changes in varve thickness due to glacial retreat, readvances, and delta progradation by using a 200-year high-pass filter (Fig. 2C). This choice was relatively conservative; very low frequency climate variability was sacrificed to ensure that the residual higher frequency

variability was relatively free of nonclimatic influences. This trend removal is similar to that used to remove nonclimatic growth-related trends in dendroclimatic records (29). The data sets were subsequently normalized to have zero mean and unit standardization. This standardization procedure is similarly conservative, in that it will tend to remove long-term trends in climatic variance, rather than impose spurious trends, because the variance is equalized in distinct components of the chronology. Of primary interest here, however, is not the overall variance or the trend therein, but the distribution of variance with regard to time scale and how it changes over time. Finally, varve thicknesses in overlapping intervals were averaged to yield one continuous sequence of 4022 years (NE varve numbers 2868 through 6900). The high correlation between the overlapping segments (30) validated averaging the constituent segments to form one continuous chronology.

We analyzed the resulting transformed varve chronology for statistically significant narrowband or harmonic signals, using multiple-taper spectral analysis (31–35), and invoked the typical climatic null hypothesis of “red noise” (34, 35) (Fig. 3). An evolutive version of the spectral analysis (using a 600-year moving window) was performed to investigate the amplitude and frequency modulation of apparent signals in the record (Fig. 4). These analyses support the existence of significant narrowband signals, with time scales remarkably similar to those observed in the modern instrumental record (36, 37, and references therein) and Holocene paleoclimate records. In particular (see Fig. 3), statistically significant variability occurs at multidecadal (>40-year) time scales, at decadal-to-bidecadal (14-to-22-year) time scales, within the conventional 2.5-to-7-year band associated with ENSO, and at the 7-to-9-year period and ~2.1-year “quasi-biennial” period. The latter two time scales may be associated with the North Atlantic Oscillation (36, 38, 39, and references therein), although the 2.1-year-period signal should be interpreted cautiously because of its proximity to the 2-year Nyquist frequency for annual sampling. The 22-year-period signal has the same frequency as the Hale solar magnetic cycle, which has been argued to have a detectable influence in modern surface temperature records (40). Only about four peaks should exceed the 99% level on the basis of chance alone (41), suggesting that most of the >99% significant peaks in the spectrum are likely indicative of climate signals with preferred time scales.

Of particular interest are the highly significant spectral peaks (>99% level) between 2.5 and 5 years, within the conventional modern ENSO bandwidth (3 to 7 years), which are highly unlikely to have arisen from chance alone (41). The concentration of peaks in 4-to-5-, 3.3-to-3.5-, and 2.5-to-2.8-year bands within

the broader ENSO bandwidth is reminiscent of the modern ENSO spectral signature (36, and references therein). The 600-year window evolutive spectrum (Fig. 4) shows significant change in the 2.5-to-5-year ENSO band over time. To ensure that these changes are not an artifact of the statistical procedure used to combine overlapping varve segments, we performed spectral analyses of individual constituent segments. These analyses reproduced the original results. Similar frequency and amplitude modulation of the ENSO signal has been suggested to occur on multidecadal and century time scales in the recent past (4, 6–8) and can arise from natural variability in the coupled ocean-atmosphere system (42). The high-frequency (2.5-to-2.8-year peak) component is relatively robust and persistent throughout the varve record from 17,500 to 13,500 cal yr B.P. In contrast, the low-frequency (3.5-to-5-year) spectrum is strong from 17,500 to ~16,200 cal yr B.P., weakened from ~16,200 to ~15,200 cal yr B.P., and diminished to absent from ~15,200 to 13,500 cal yr B.P. This long-term trend of continual weakening of the interannual variability throughout the 4000-year record appears to be systematic rather than random in nature and may be associated with low-frequency astronomical forcing of ENSO (15, 16). The high-frequency (2.5-to-3-year) component of ENSO is typically associated with the lifetime of ENSO events and the El Niño/La Niña alteration, whereas the lower frequency (3-to-7-year) component is typically associated with the spacing of large events (43, 44). The absence of the lower frequency component of variability during the latter part of the interval studied here (i.e., the latest Pleistocene) is consistent with the theoretical prediction (15, 16) of fewer large events in the early Holocene.

Only the 15,000-year Ecuador debris-flow record of ENSO variability (13) overlaps in time with the NE varve chronology. These sediments show a similar trend of weakened ENSO-band variability from 15,000 to 7000 cal yr B.P. However, the lack of annual laminations in that record limits interpretations regarding true interannual ENSO variability. Rather, this record exhibits a dominant decadal (~15-year) time scale, which is typically associated with extratropical feedbacks (6, 45, 46) rather than the tropical dynamics governing interannual ENSO variability. Their observations in fact are consistent, for example, with the apparent pattern during the 17th through early 18th century, during which interannual variability in ENSO indices appears to have been largely suppressed, whereas decadal-scale variability nonetheless persisted (8).

Our results are consistent with model predictions using early Holocene insolation parameters (15). The coupled ocean-atmosphere model results for 11,000 cal yr B.P. (15) suggest a decreased amplitude of low-frequency (e.g., 4-to-7-year time scale) inter-

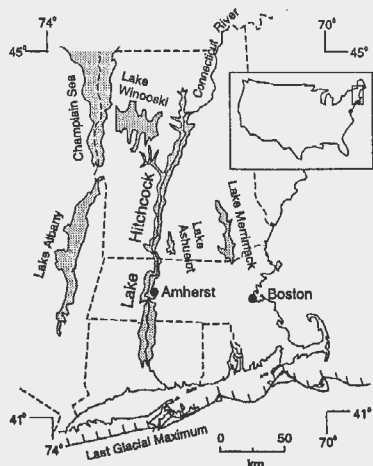


Fig. 1. Location of Glacial Lake Hitchcock in New England and nearby contemporaneous glacial lake systems from which the NE varve chronology was created.

annual variability but a persistence of high-frequency (e.g., 2-to-3-year range) ENSO variability. Lui *et al.* (15) suggested that this pattern of reduced low-frequency ENSO variability was caused by Asian monsoon intensification during increased insolation in the early Holocene. The gradual increase in solar insolation during the late Pleistocene may have caused the decreased low-frequency (3.3-to-5-year) ENSO variability seen in the NE varve chronology from ~15,500 to 13,500 cal yr B.P.

Despite the similarities between the ENSO pattern seen in the NE varve chronology and modern and paleoclimatic records, it is not possible to determine whether the changing intensity of the ENSO-like signal is directly related to El Niño intensity in the tropical Pacific Ocean. As discussed earlier [see (20, 21)], one would expect ENSO to influence seasonal melt patterns in New England during the late Pleistocene. However, ENSO teleconnections into New England may have been continually altered as boundary conditions changed during ice retreat. It is plausible that the retreat of the Laurentide ice margin, with its tendency to

deflect extratropical storm tracks, could have led to a slow northward migration of storm tracks and thus ENSO teleconnections into North America, which represent a modulation of these storm track patterns. However, varve deposition in Glacial Lake Hitchcock followed the ice margin with progressively younger

varves deposited northward, therefore limiting this effect. Although the persistent high-frequency component of the ENSO band throughout the NE varve chronology argues against any dominant influence of nonstationary teleconnections, the possibility nonetheless remains that the observed changes are in u-

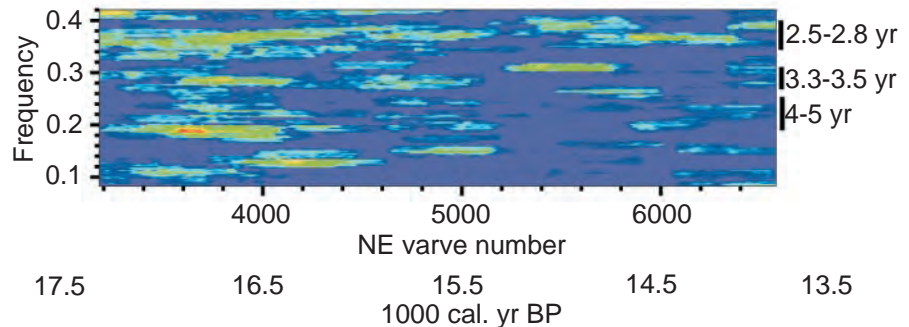


Fig. 4. Evolutive spectral analysis from the NE varve chronology data set (NE varve numbers 2868 through 6900) covering the interval from 17,500 to 13,500 cal yr B.P. The spectrum is normalized everywhere by the median red noise level, so that the amplitude shown is proportional to the associated level of statistical significance. The background color is set to the median red noise level (50% confidence), and features evident above the background are significant at the 50% confidence level, with the lightest features being the most significant.

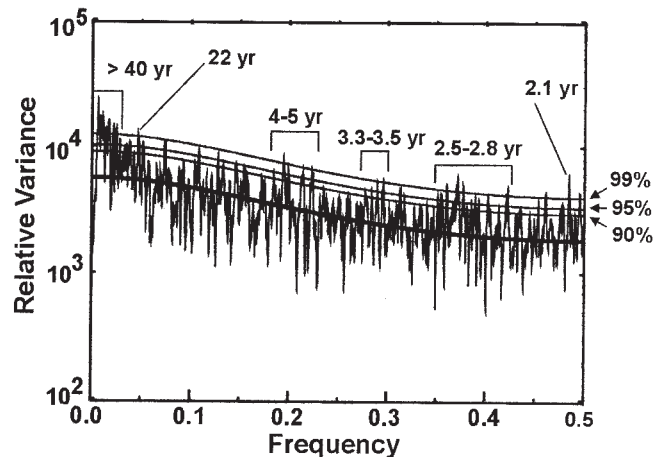
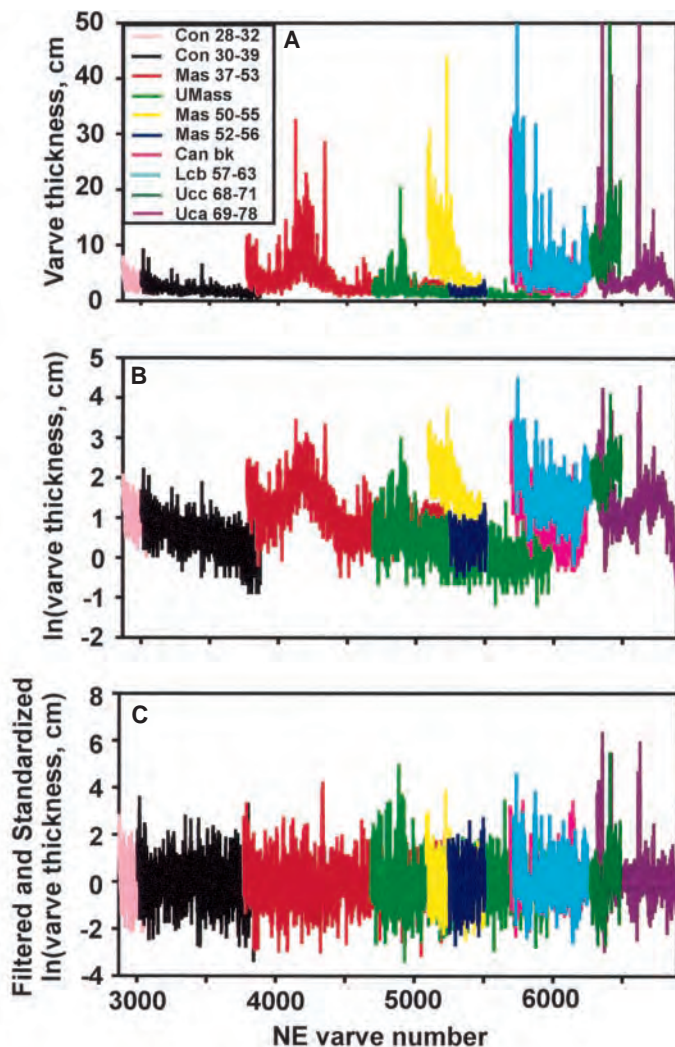


Fig. 2 (left). Normal curve varve thickness data sets from the Glacial Lake Hitchcock portion of the NE varve chronology used for spectral analysis (NE varves 2868 through 6900, 17,500 through 13,500 cal yr B.P.) (data sets used are available at www.sciencemag.org/feature/data/1047305.shl). (A) Original raw varve couplet thickness measurements, (B) natural log of varve thickness data, and (C) data after standardization (zero mean and unit standard deviation) and high-pass filtering (removal of time scales of 200-year and longer periods) are given. Fig. 2 (right). Multitaper method spectral analysis (1, 33) of NE varve chronology data set (NE varve numbers 2868 through 6900) covering the interval from 17,500 to 13,500 cal yr B.P. The median red noise level is shown by the thick curve, and 90, 95, and 99% confidence levels relative to red noise are shown by the thin curves. Significant spectral peaks are labeled.

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enced by some combination of changes in the nature of ENSO teleconnections into North America and changes in ENSO itself. Our primary conclusion, that ENSO was active during the late Pleistocene, follows, however, in either scenario.

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Reduction of Tropical Cloudiness by Soot

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Measurements and models show that enhanced aerosol concentrations can augment cloud albedo not only by increasing total droplet cross-sectional area, but also by reducing precipitation and thereby increasing cloud water content and cloud coverage. Aerosol pollution is expected to exert a net cooling influence on the global climate through these conventional mechanisms. Here, we demonstrate an opposite mechanism through which aerosols can reduce cloud cover and thus significantly offset aerosol-induced radiative cooling at the top of the atmosphere on a regional scale. In model simulations, the daytime clearing of trade cumulus is hastened and intensified by solar heating in dark haze (as found over much of the northern Indian Ocean during the northeast monsoon).

A primary objective of the Indian Ocean Experiment (INDOEX) was to quantify the indirect effect of aerosols on climate through their effects on clouds (1). Conventionally, increased aerosol concentrations are expected to increase cloud droplet concentrations, and hence, total droplet cross-sectional area, thereby causing more sunlight to be reflected to space (2). Furthermore, model simulations

of marine stratocumulus (3–5) and observations of ship tracks (6–8) suggest that increased aerosol concentrations can enhance cloud water content, physical thickness, and areal coverage by decreasing precipitation. Deep layers of dark (solar-absorbing) haze were observed over much of the tropical northern Indian Ocean in February–March of 1998 and 1999 during INDOEX (9, 10). The clouds observed in the Northern Hemisphere were typically embedded in the haze (Fig. 1). In contrast to the conventional expectation that aerosols augment cloud depth and coverage, very sparse cloud cover is found in that region during that time of year (11). These INDOEX observations suggest a new mechanism by which aerosols impact clouds, in which a dark haze can significantly reduce areal coverage of trade cumulus (the predom-

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Lake Hitchcock, fed by glacial melt water for approximately 3,000 years, disappeared from the New England landscape about 12,000 years ago. Glacial lakes such as Lake Hitchcock formed geological deposits known as varves, each varve is composed of a couplet consisting of a whitish-gray clay layer and a yellowish-brown silty/sandy layer; a couplet represents a single year, the silty/sand deposited in the short summer during ice melt and the clay layer deposited in the long winter when the lake was frozen.

Although Lake Hitchcock disappeared before the first Amerindians entered the Connecticut River Valley, divers can still explore its waters; and that is what we were doing, forty feet under the Connecticut River, exploring a lake that vanished. The river's current had torn a small underwater canyon through a large block of lake sediment. We swam gently down this cleft, the orderly varves were a time warp back to the ice ages. As we descended deeper and deeper, year after year, decade after decade, and, finally, forgotten centuries passed and were lost from view. What events did they chronicle? The birth of a mastodon, the roar of a saber tooth cat, or perhaps something more prosaic but certainly more important -- the survival of the first tree seedling after the ice retreat.

At 60 feet the varve layers abruptly came to an end. With wetsuits streaked with gray glacial clay, we knelt on the bottom and looked up. Surrounding us was an underwater amphitheater of varves with the upper layers disappearing into darkness. Stacks of clay occasionally crumbled and fell, leaving a plume of gray "smoke" as the clay particles dissolved into the water. Bass and other fish were attracted to these small underwater avalanches in hopes of catching the animals inhabiting the varves.

Lake Hitchcock was one of the largest of the glacial lakes in New England. It stretched from mid-Connecticut to northern Vermont, approximately 175 miles. The impoundment resulted from glacial deposits at Rocky Hill, Connecticut that dammed the ice melt as the last glacier retreated northward. The Rocky Hill dam was breached and the lake drained about 12,000 years ago. The Connecticut River generally follows the course of Lake Hitchcock and, in many places, the river has eroded into and sometimes through the lake-bottom sediments. Diving in these portions of the river is, in many respects, like going back in time and exploring the bottom of Lake Hitchcock.

Descending through the river's waters to the lake sediments, the first varves often appear as ghostly white sheets of clay. Closer examination of these clay surfaces reveals a Swiss-cheese-like texture caused by countless burrows of chironomid larvae, an as yet

unknown species of the genus *Axarus*. Chironomids are midges (insects) whose larvae are aquatic, contain hemoglobin, and are an important component of freshwater food webs. Breaking up a piece of clay releases these red worm-like larvae.

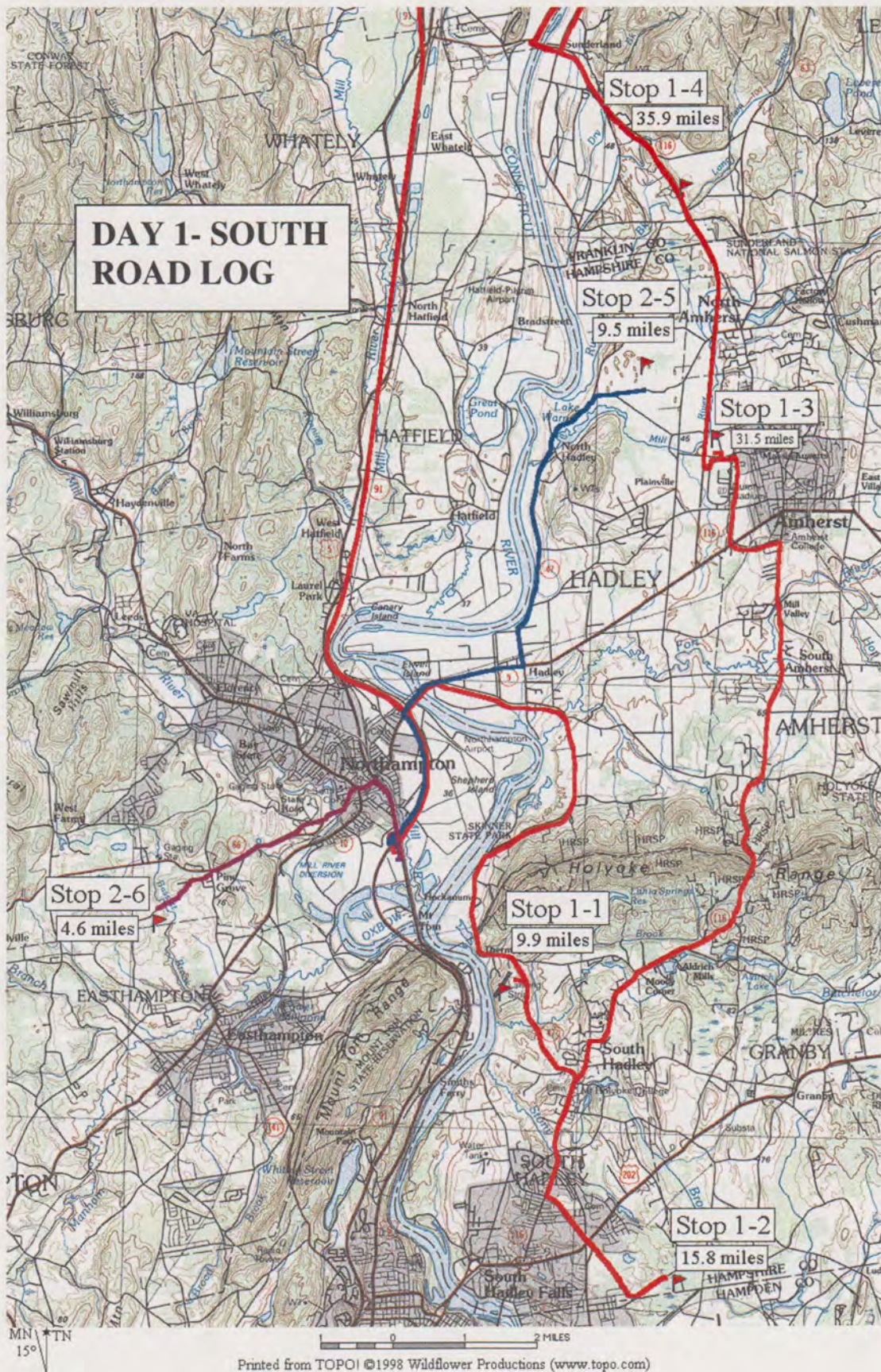
In the deeper part of the river, erosion has cut into the varves and they can be viewed in cross-section. Often the sandy/silty layer of a varve couplet is eroded and undercuts the clay layer. Looking carefully with a dive light into these crevices may reveal a pair of antennae sensing the environment. Looking closer, the diver will find a pair of stalked compound eyes looking back! The Connecticut River crayfish (*Orconectes limosus*), a solitary bottom dweller, spends its day hiding in crevices or under stones in the river. Its food includes insect larvae (chironomids), smaller crustaceans and dead animal matter. Crayfish are the food of choice for the many carnivorous fish in the river, thus this crustacean's reclusive nature is not surprising.

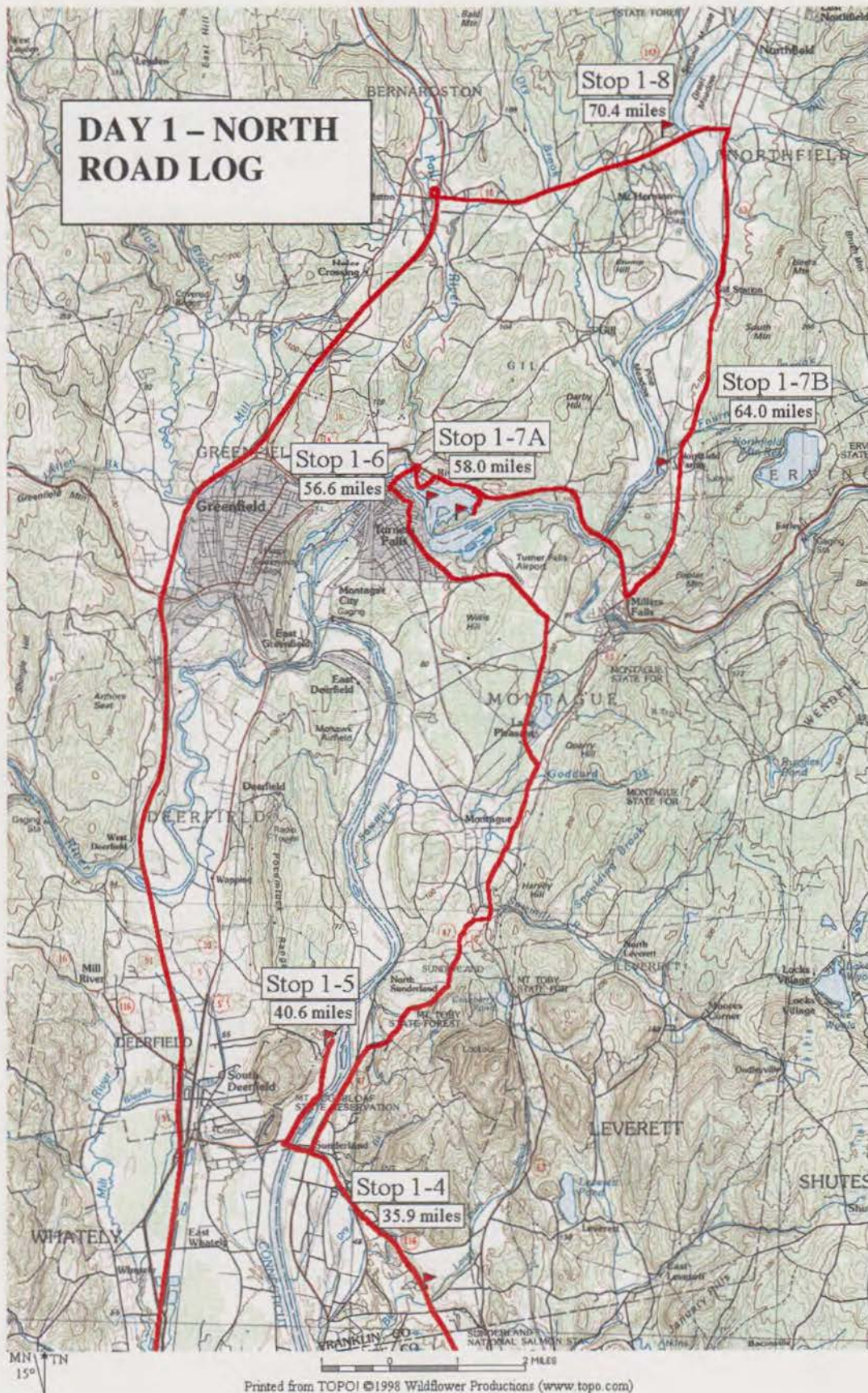
Very curious concretions of organic matter and carbonate are embedded in the clay portions of the varves and literally carpet the river bottom downstream of exposed and eroding Lake Hitchcock sediments. These concretions are tabular, approximately 6-8 inches across, and resemble puzzle pieces with circular holes punched in them; no two are identical and almost all are very attractive with sensuous, sinusoidal curves. All Lake Hitchcock divers return with a handful as mementos of their visit.

Each varve couplet marks a year; a continuous series of varves is the geological equivalent to the annual growth rings of a very old tree. The thickness of the couplet layers is a record of past environments. The varve couplets beneath the Connecticut River have never been studied and thus may offer new information concerning the rate of ice retreat and climate changes as we exited the last ice age.

In the Northeast the glacier began to retreat about 21,200 years ago. In its wake a series of freshwater glacial lakes were formed. The largest of these was glacial Lake Connecticut, which later became Long Island Sound when sea level rose. Two glacial lakes are traversed by the Hudson River: Lake Hudson in the south and Lake Albany in the north. In New Hampshire the Merrimack River follows the bed of glacial Lake Merrimack. In all of these sites, divers should discover ice age lake sediments similar to those under the Connecticut River.

**Check out the Connecticut Valley Web Site organized by Ed at
<http://www.bio.umass.edu/biology/conn.river/>**





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DAY 1: FIELD TRIP STOPS AND ROAD LOG

From Northampton, MA, at I-91 exit 18, head north to exit 19 (2.38 miles). Take Route 9 east (right) over the Coolidge Bridge . Just after the Bay Road Stay on Rt 47 for 5 miles. Go Right into Dry Brook Hill gravel pit. (odometer 9.89 miles from the Inn at Northampton, hereafter cumulative miles).

STOP 1-1. DRY BROOK DELTA, South Hadley, MA (Mt. Holyoke quadrangle) ; Janet Stone and Al Werner.

Dry Brook Hill in South Hadley, Massachusetts is an ice-marginal delta built into glacial Lake Hitchcock at a time when the ice margin stood along the north and northwest side of the Holyoke Range and a narrow tongue of ice protruded through the gap (see fig. 25, Donner, 1991; Saines, 1973).

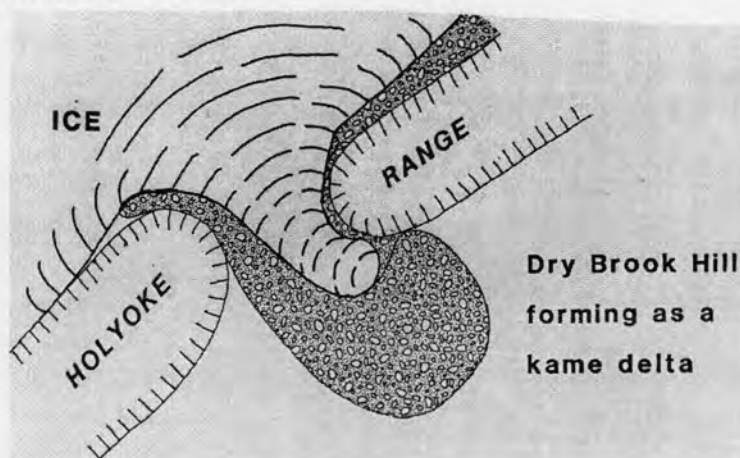
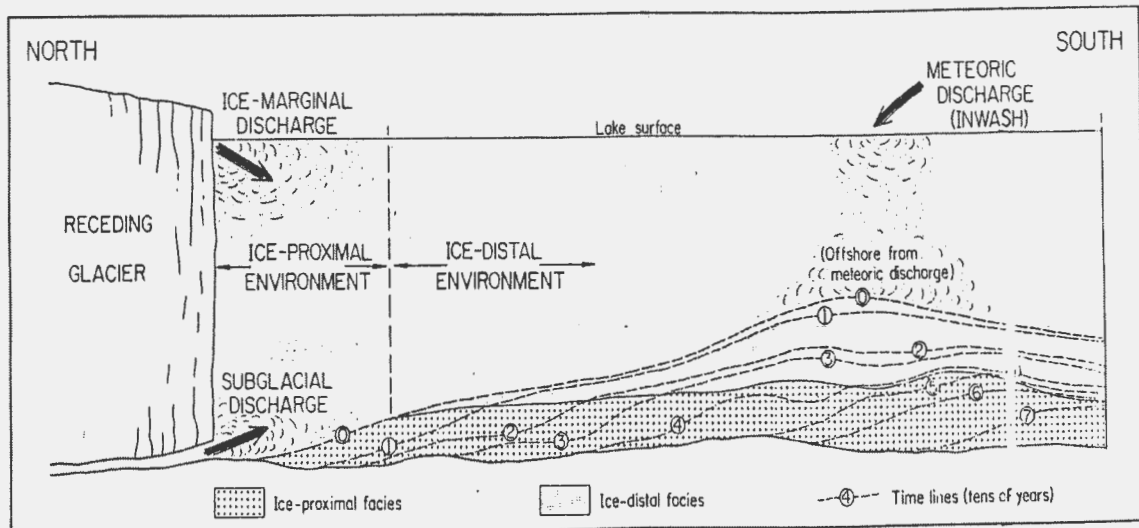


Figure 1-1A. Reconstruction during deglaciation.

Deposition of this landform began as sediment-laden meltwater (under hydrostatic head) issued from a subglacial tunnel onto the floor of Lake Hitchcock and constructed a subaqueous lacustrine fan. Deposition continued after fan sediments had built up to lake level and the final construction of the landform was accomplished by delta progradation southeasterly into the lake. Interpretation of numerous ground-water monitoring well logs and 3-D reconstruction of the distribution of sedimentary facies within Dry Brook Hill provide the evidence that the deltaic landform is cored by lacustrine fan beds in the northwest and central parts, although these areas have not been exposed. The northwest side of Dry Brook Hill is underlain by coarse, poorly sorted gravel and sand which extends beneath glacial lake clays in the valley of Dry Brook at the northwest base of the hill. These gravels are a highly productive confined aquifer tapped by a South Hadley public-supply well, located at the mouth of Dry Brook (Delude, 1995).



from Ridge and Larsen, 1990.

Figure 1-1B. Depositional systems possibly operating in the vicinity of Dry Brook Hill, taken from Ridge and Larsen, 1990, GSA paper.

The eastern and southern sides of Dry Brook Hill are the frontal foreset slopes of the delta, underlain by sands and clays. The steep western slope of the hill is an erosional scarp, cut first by water spilling from Lake Hitchcock north of the Holyoke Range, after the southern basin had drained (Stone and Ashley, 1995; Stone and others, 1998), and later by the ancestral Connecticut River as it continued to incise the entirely drained lake bed. A small remnant of the delta remains on the west side of the river (245-ft hill on west side of I-91). The Dry Brook delta plain has a surface altitude of the of 265-270 ft, and the Stop 1 gravel pit (owned by Mr. Leo Ouelette, of South Hadley) exposes 10-12 ft (3-4 m) of topset gravel and sands and gravel overlying about 60 ft (20 m) of easterly dipping foreset sands. The topset-foreset contact in this pit was measured at 255 ft (77.7 m) altitude (Jacobson, 1981). This altitude is 5-6 ft (1.5-2 m) higher than the projected stable level (249.6 ft (76.1 m), Koteff and Larsen, 1989) of Lake Hitchcock at this latitude, indicating that perhaps the lake level had not quite reached stable stage at the time of its deposition (see discussion of Lake Hitchcock high-level, stable-level and post-stable-level stages in Koteff and others, 1988).

Pearl City Delta: The Pearl City delta lies about 2 miles to the east of Dry Brook Hill. This deposit is a distal meltwater-fed delta consisting of fluviodeltaic sediments laid down by meltwater flowing westward in the Bachelor Brook valley, originating from ice margin positions on the east side of the basin, north of the Holyoke Range. The delta-plain surface is at 255 ft in altitude, rising eastward, up valley, to 265 ft. (see attached figure). Several gravel pits excavated into the 255-ft surface have revealed topset-foreset contacts; two altitudes are reported by Koteff and Larsen (1989)-- an altitude of 247 ft

(75.3 m) from Jahns and Willard (1942) and 252.8 ft (77.1 m) measured by Fred Larsen during their study. An excellent current exposure along Rt. 116 near the west end of the delta plain reveals about 6-8 ft (2-2.5 m) of sandy topset beds overlying 50-70 ft (15-20 m) of west-dipping foreset beds. Unfortunately we were not able to obtain access for the fieldtrip to visit this exposure; however, the "eyeballed" level recorded by the topset-foreset contact appears to fall near the projected stable level of 249.6 ft (76.1 m). The Pearl City delta built into Lake Hitchcock concurrently with or slightly later than the Dry Brook delta; its long narrow surface, sloping westward from 270-250 ft, and 6-ft range of topset-foreset altitudes may indicate that it was constructed during continued lowering of lake level. The foreset slope of the delta drops fairly steeply from 250 ft to its base at about 215 ft; the long sandy plain sloping westward from 215 ft to 185 ft along Pearl Street was likely constructed as Lake Hitchcock lowered some 30-40 ft (10-12 m) from its stable level.

Alluvial Fan: Another feature in this vicinity (and at a similar altitude) related to lowering of Lake Hitchcock from its stable level is the alluvial-fan shaped landform sloping west from the South Hadley Commons area (see attached figure). Recent construction exposures in the Priestly Estates subdivision indicate that thin (2-3 m) alluvial sand overlies thinly bedded varved clay. Ripple cross-laminae and climbing ripples in the sand indicate westward transport and suggest rapid sedimentation. Deposition of the fan is thought to be associated with the lowering of Glacial Lake Hitchcock prior to incision of the present Stony Creek valley.

The question to be considered here concerning the history of lake-drainage history is whether evidence points to a continuum of deposition from the stable-level features (requiring meltwater input) to the post-stable level features, or to a long hiatus in time followed by deposition of the post-stable features by meteoric water.

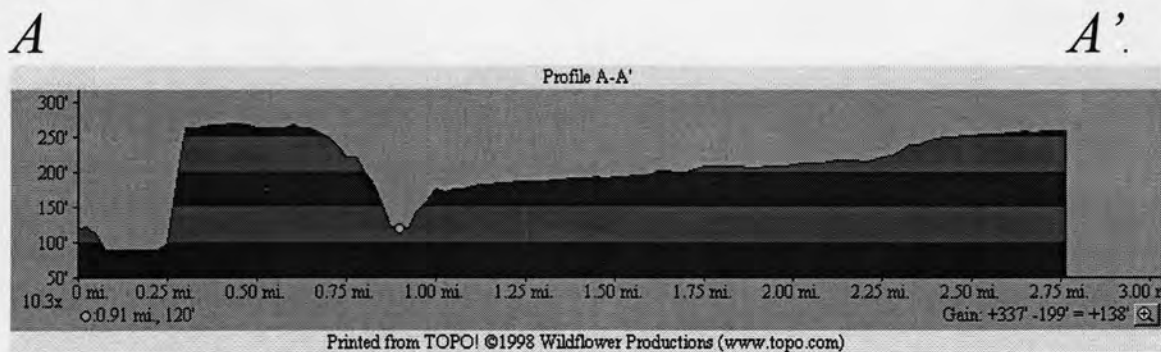


Figure 1-1C. Cross Section of Dry Brook and Pearl City Deltas. Location on map on next page.



Figure 1-1D. Location of the cross section shown on the previous page highlighting the elevation differences between the Dry Brook and Pearl City Deltas.

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Delude, N.A., 1995, The hydrogeology of Dry Brook Aquifer, South Hadley, Massachusetts; unpublished Honors thesis, Mount Holyoke College, South Hadley, Massachusetts, 48 p.

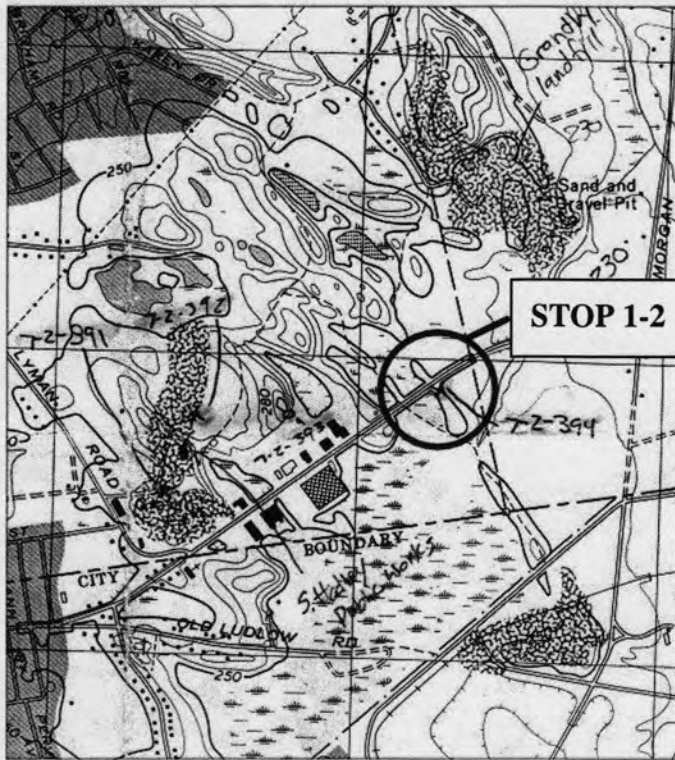


Figure 1-2A.
Topographic map of
the site location.
Contour interval is
10 feet, Springfield
North Quadrangle.

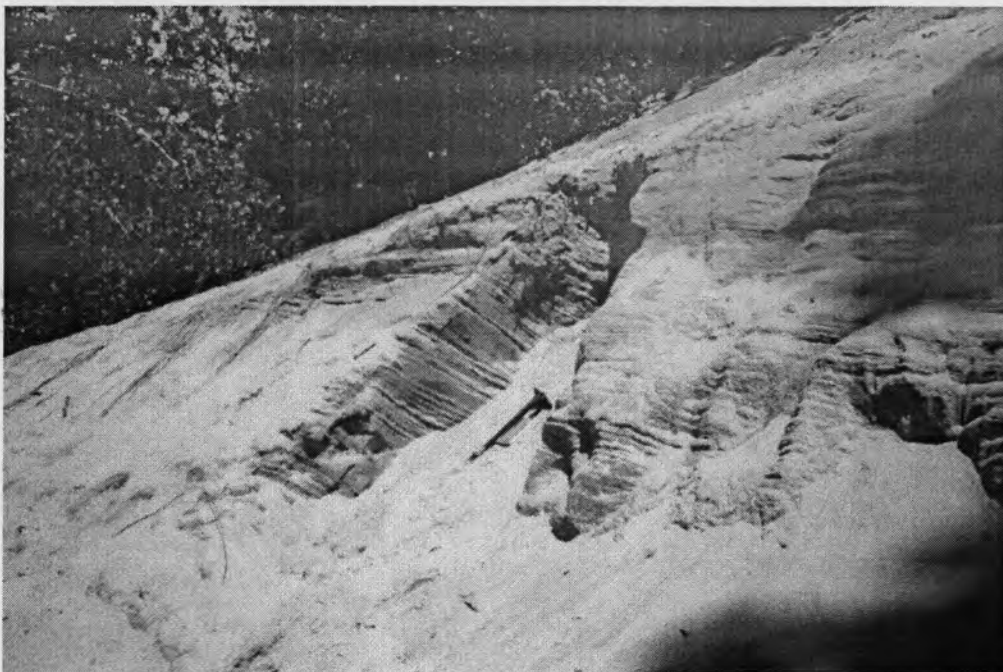


Figure 1-2B. Photograph of the South Hadley transverse dune clearly indicating the characteristic dipping foreset beds. Shovel for scale

- Donner, J.E., 1991, Origin, Extent and Recharge of Dry Brook Aquifer, South Hadley, Massachusetts: unpublished Honors thesis, Mount Holyoke College, South Hadley, Massachusetts, 42 p.
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- Saines, Marvin, 1973, Hydrogeology and Hydrogeochemistry of parts of South Hadley, Massachusetts: unpublished Ph.D. thesis, Department of Geology and Geography, University of Massachusetts, Amherst, Mass., 376 p.
- Stone, J.R. And Ashley, G.M., 1995, Timing and mechanisms of glacial Lake Hitchcock drainage: Geological Society of America Abstracts with Programs, v. 27, no. 1, p. 85.
- Stone, J.R., Schafer, J.P., Stone, J.R., Schafer, J.P., London, E.H., DiGiacomo-Cohen, M., Lewis, R.S., and Thompson, W.B., 1998, Quaternary geologic map of Connecticut and Long Island Sound Basin: U.S. Geological Survey Open-file Report 98-371, scale 1:175,000, with text, 69 p.

Roadlog Cont. From Dry Brook Hill. Go South 1.9 miles to S.Hadley Center on Rt. 47. At the town common, Rt 47 merges with Rt 116. Follow Rt. 116 south for 1.9 miles. Take a left on Rt 33 eastward, go 0.8 miles and cross Rt 202; continue and take immediate left onto Old Lyman Road. Go to the end of the road (about 1 mile). Take left onto New Ludlow Road and go 0.6 miles to sand dune exposure on SE side of road. (cumulative miles 15.8 miles)

STOP 1-2 SOUTH HADLEY SAND DUNE - Road cut SE of South Hadley along New Ludlow Road ~702200m E, 4676900m N UTM, 250 ft asl (76 m asl) Springfield North Quadrangle ---Tammy Rittenour

Exposed in the road cut is a sand dune that formed on the surface of the Chicopee Delta. The linear morphology of this sand dune (Figure 1-2A) and the presence of steeply dipping (~29°) grain fall foreset beds (Figure 1-2B) indicates that this dune was an actively migrating transverse sand dune. Foreset dip directions of N 75° E to N 80° E and the linear trend of the dune indicate a westerly to southwesterly wind direction during formation.

This transverse dune is part of a small dune field located to the north and west of this site (Figure 1-2A). The dune field is composed primarily of transverse and compound dune forms. Sand sheets surrounding the dune field and are composed of horizontal, inversely-graded beds formed by translational ripples. In places basal foresets from dunes that had migrated through the sand sheets are truncated by horizontal laminae. At one location wind-polished pebbles (ventifacts) were found along the erosional contact between the basal foresets and overlying translational ripple strata.

Optically-stimulated luminescence (OSL) age estimates were obtained from the transverse dune at this site and other dunes in the Glacial Lake Hitchcock basin (OSL analysis by Steve Forman at the University of Illinois, Chicago). OSL dating measures the length of time since sediments were last exposed to sunlight. The luminescence signal, which is removed during sediment transport and exposure to light, accumulates in mineral grains (quartz was used here) during exposure to ambient radioactive isotopes in the surrounding sediment. The amount of luminescence acquired by the sediment is related to the length of burial and the chemistry of the surrounding sediment (U, Th and K content).

OSL age estimates from sand dunes located on a number of genetic surfaces within the Connecticut River valley in Massachusetts indicate that most of the dunes formed between 14.0 ± 1.0 and 14.4 ± 1.0 cal. kyr BP (see table of ages below). This translates to 12.0 to 12.4 ^{14}C kyr BP. The exceptions to this are the Turners Falls dune located on the Montague delta which gave an OSL age of 12.1 ± 1.4 cal. kyr BP (~ 10.3 ^{14}C kyr BP) – possibly due to reactivation of the dune during the Younger Dryas. The OSL sample taken from sand 65 cm above the till contact at the Hillside Plastic Company site sample # 1 produced an age of 12.3 ± 2.0 cal kyr BP (10.4 ^{14}C kyr BP) while sample #2 which is stratigraphically higher, produced an age of 14.0 ± 1.0 cal kyr BP (12.0 ^{14}C kyr BP). This age difference may have been due to water saturation of the lower sample just above the till contact (sample moisture attenuates the OSL signal by blocking radioactive alpha, beta and gamma particle from reaching the mineral grains). The third suspect OSL age estimate came from the South Hadley dune (dune exposed at this stop). Steve Forman reported that the OSL signal from this dune sample had unusual scatter in the OSL results and he was only able to confine the age of this dune to less than 20.0 ± 2.0 cal. kyr BP (~ 17.0 ^{14}C kyr BP). However, it is possible that this dune formed prior to the dunes north of the Holyoke Range because the Chicopee Delta would have been available for eolian deposition before the till, delta, terrace and lake bottom surfaces to the north. Despite these suspect OSL age estimates, the OSL age results suggest that dune formation occurred between 14.0 ± 1.0 and 14.4 ± 1.0 cal. kyr BP.

OSL analysis performed by Steve Forman, University of Illinois Chicago. Green light stimulation was completed on the 100 to 150 μ fraction and the paleodose was determined by the multiple aliquot additive dose technique. Moisture content of 10

$\pm 3\%$ was used for final age calculation on all of the samples except Hillside Plastics sample #1 where $50 \pm 10\%$ moisture was used. The Aitken, 1998 dose rate calibration data were used to determine the OSL age from the sediment chemistry.

<u>OSL Results</u>	<u>OSL age, cal. kyr BP</u>
<u>Turners Falls Dune</u>	12.1 ± 1.4
Dune on the Montague Delta	
Transverse dune (westerly wind)	
<u>Hillside Plastic Company Sand # 1</u>	12.3 ± 2.0
Eolian sand overlying till	
65 cm above till/sand contact	
<u>Hillside Plastic Company Sand # 2</u>	14.0 ± 1.0
Eolian sand overlying till	
2.5 m above till/sand contact	
<u>Hadley Sand Dune</u>	14.3 ± 1.6
Dune on exposed lake bottom in the Hadley basin	
Transverse dune (westerly wind)	
<u>Montague Dune</u>	14.4 ± 1.0
Dune on an early terrace graded to Lake Hadley	
Echo dune at base of delta (westerly wind)	
<u>South Hadley Sand Dune</u>	$<20.0 \pm 1.9$
Dune on the Chicopee Delta	
Transverse dune (westerly wind)	

REFERENCE FOR STOP 2

Aitken, M.J., 1998, An Introduction to Optical Dating: Oxford Science Publications, Oxford U.K.

RoadLog continues. Return to Rte 116. Take a right (cumulative miles = 18.2) and continue north on Rte 116 over the Holyoke Range to Amherst (distance of 11.2 miles, Amherst College on your right). Take a left at the light on Rte 9 (going west) and continue about 1 mile to University Drive. Take a right onto University Drive going north and continue to t-intersection with Memorial Drive. Take a left go a few hundred yards and take a right onto Mullins Way and continue north 0.3 miles to UMass National Engineering site. (Cumulative = 31.5 miles)

STOP 1-3A UMASS CAMPUS CORE SITE AND RESULTS

University of Massachusetts, Amherst National Engineering test site; ~703000m E,
4696000m N UTM, 140 ft asl (43 m asl); Mt Toby Quadrangle --Tammy Rittenour
and Julie Brigham-Grette

OVERVIEW OF UMASS CORE RESULTS

- Two cores were retrieved from the UMass campus, one was drilled to bedrock (32 m) and a second shorter core was taken to ensure complete recovery of the upper portion of the varve sequence (7.6 m).
- The UMass core covers NE varve 4638 – 6027, a total of 1,389 varves.
- A radiocarbon date from small plant fragments picked from varves in a contorted zone between NE varve 5761 – 5768 produced a $\delta^{13}\text{C}$ corrected AMS age estimate of $12,370 \pm 120$ ^{14}C yr BP ($14.3^{+1.2}_{-.0.4}$ cal kyr BP).
- Age estimates from the top and bottom of the UMass core indicates that it covers varve deposition from 15.4 to 14.0 cal. kyr BP (12.8 to 12.0 ^{14}C kyr BP). <- note that there is a large radiocarbon plateau during this time (~1400 cal yrs = only 800 radiocarbon years).
- Glacial Lake Hitchcock completely drained in the Hadley basin sometime after NE varve 6027 (last varve deposited at top of core) at 14.0 cal. kyr BP (12.0 ^{14}C kyr BP).
- The ice margin had retreated from the Amherst, MA, and the first ice-proximal varve was deposited (NE varve 4638) at 15.4 cal. kyr BP (12.8 ^{14}C kyr BP).

CORE DESCRIPTION

In the fall of 1997, two 10-cm diameter cores were drilled with an auger-driven truck-mounted drill rig on the University of Massachusetts, Amherst campus. The core site is located 1.5 km from the shore of Glacial Lake Hitchcock on a 43 m asl lake-bottom surface in the broad, deep Hadley basin. In this area the lake was approximately 20 km wide and the stable lake level was at 90 m asl. Water depth at the core site was 77.5 m at the initiation of varve deposition.

The longest core, 32 m, was drilled to bedrock collecting a continuous sequence of varves ranging from thick (7 – 55 cm) ice-proximal varves at the base of the core, up through varves of average thickness for the region (1 – 4 cm), and into thin varves with relatively coarse-grained summer layers and thinner winter clay layers (Figure 1-3A). The second core, 7.6 m, was obtained to ensure complete coverage of the upper portion of the varve section and the transition to non-varved lacustrine silt and sand. The cores were obtained in 5 ft drives with minimal, 0 – 2 cm, sediment lost between drives.

Sediment disturbance from the coring process only occurred just above the bedrock contact where the varves were thick, coarse-grained and saturated with ground water.

Varve Description

The varves are typical clastic glacial varves formed due to seasonal changes in sediment input (melt season vs. lake ice cover). The varve couplets are composed of coarser summer layers that grade into winter clay layers. Varve deposition conforms to Kuenen's (1951) model of varve sedimentation by density flows (under-, inter- and overflows depending on meltwater and lakewater density) during the summer melt-season and settling of clay during the quiescence of ice cover. This has been demonstrated by the presence of multiple graded beds within the melt-season layer, uniformity of clay layer thickness and sediment distribution throughout the lake basin (Ashley, 1972).

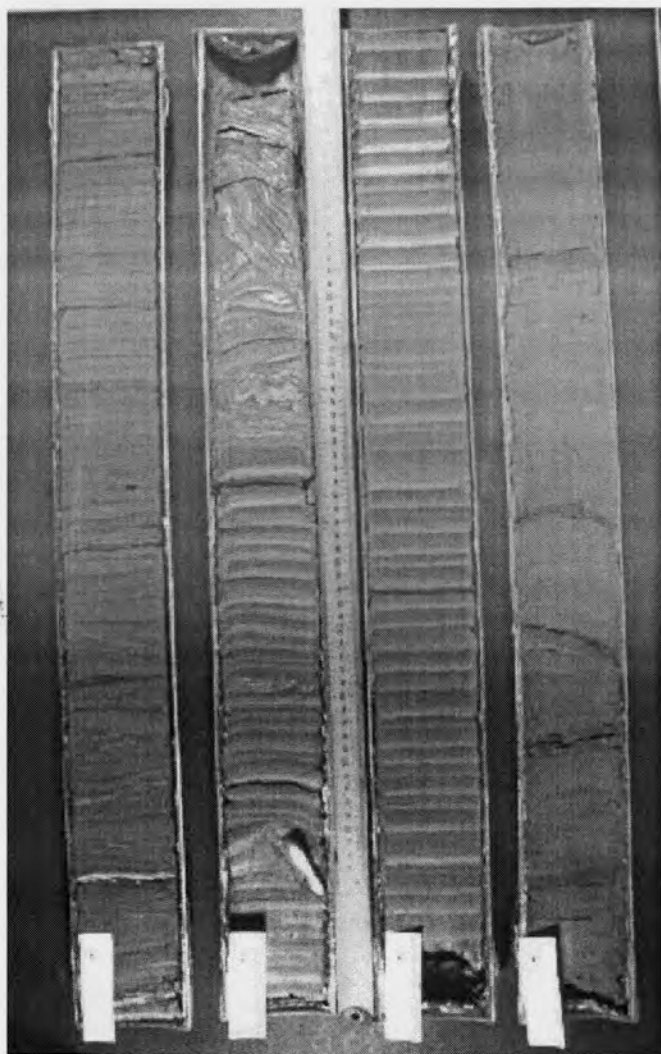


Figure 1-3A. Picture of sections of the UMass core. From left to right, the first core section is from 7.5 to 10 ft (2.3 – 3 m) depth, the second core is from 12.5 to 15 ft (3.8 – 4.6 m) depth, the third core is from 42.5 to 45 ft (12.9 – 13.8 m) depth, and the fourth core was from near the base of the core at 87.5 to 90 ft (26.7 – 27.4 m) depth. The varve deposition proceeded from thick ice-proximal varves at the base of the core up to varves of moderate thickness for the region and into thin silty varves at the top of the core. The second core from the left contains syndepositional contorted slump sediments and a postdepositional carbonate concretion which was tilted during coring. A meter stick with marks in centimeters is provided for scale

For each varve, the total varve couplet thickness, summer silt layer thickness and winter clay layer thickness was measured with calipers and recorded with a short sedimentological description. Once the measurements were completed

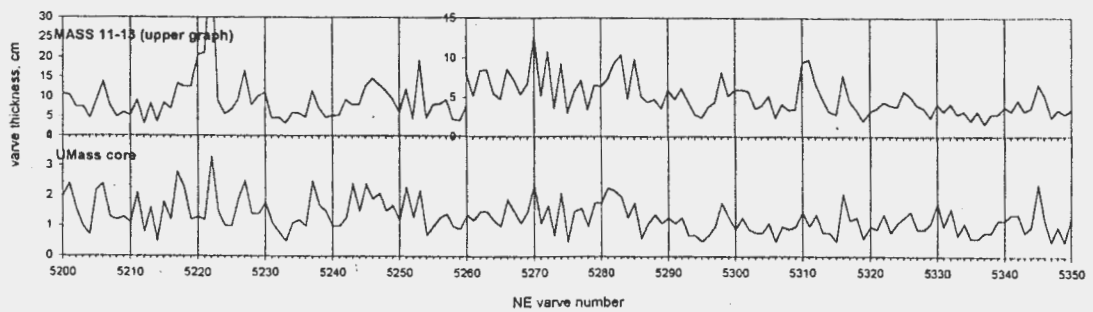
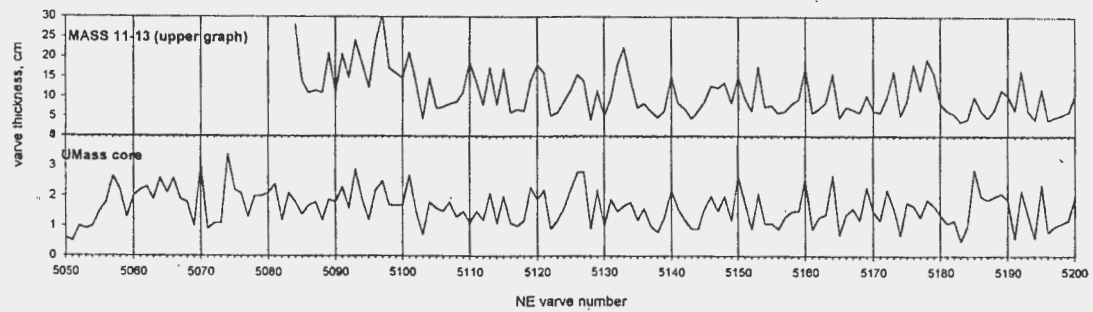
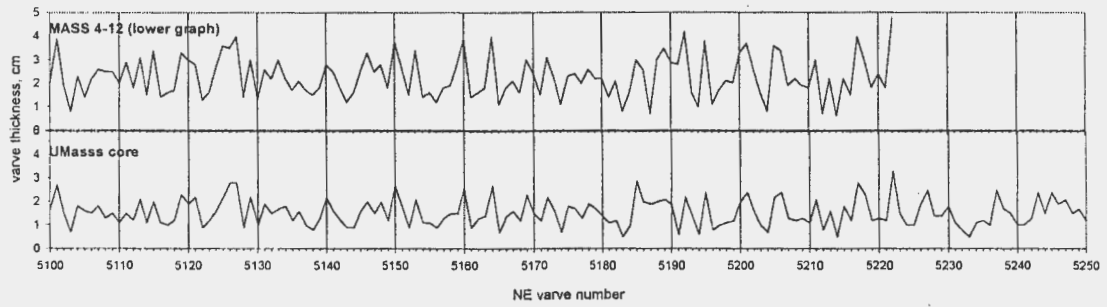
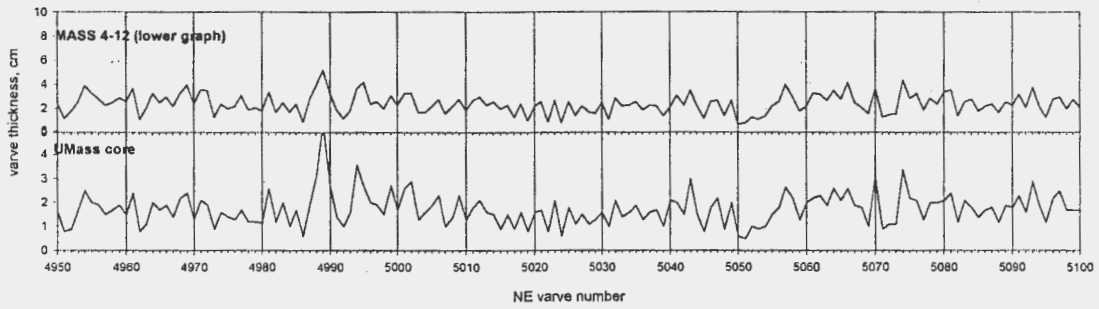
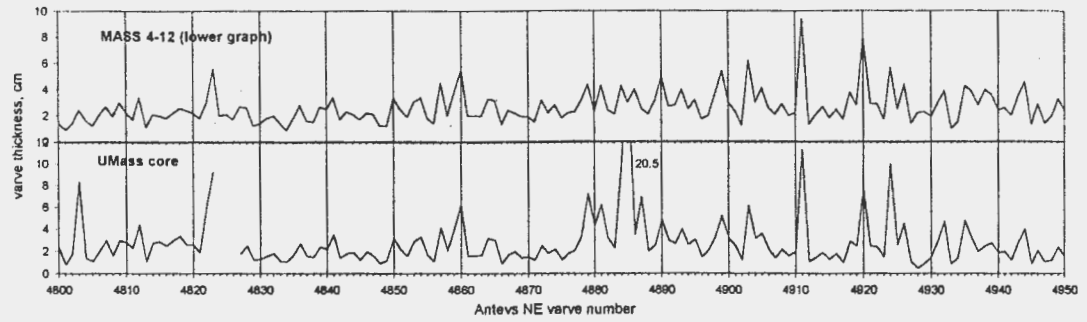
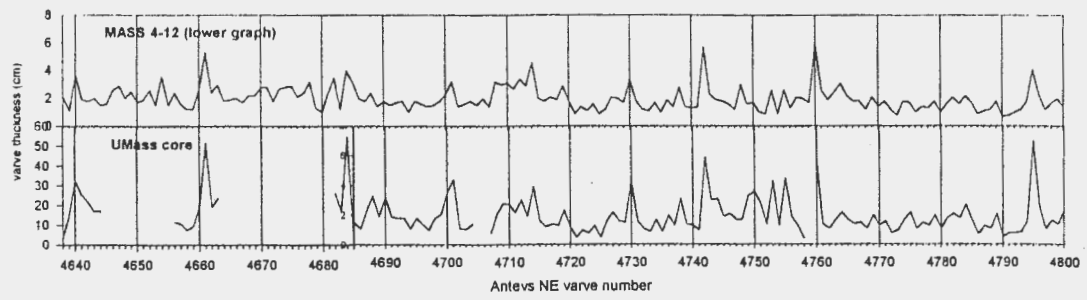
varve couplet thickness plots were compiled. Excellent varve for varve correlation between the two cores was obtained by matching varve couplet thickness plots for the overlap section. The two cores were then combined to form one continuous 33.5 m (110 ft) series of varves.

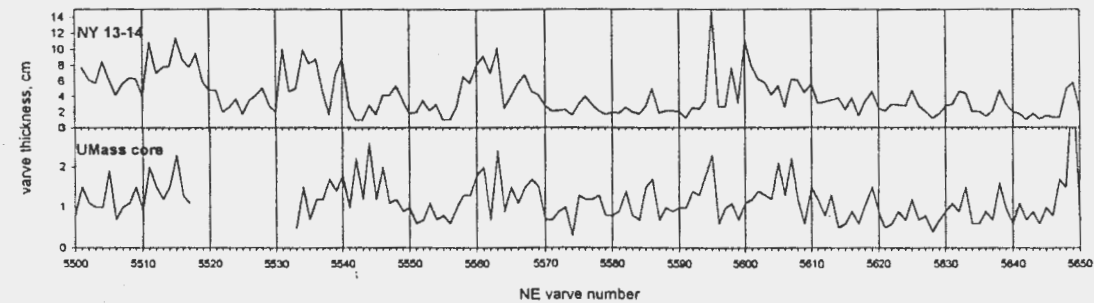
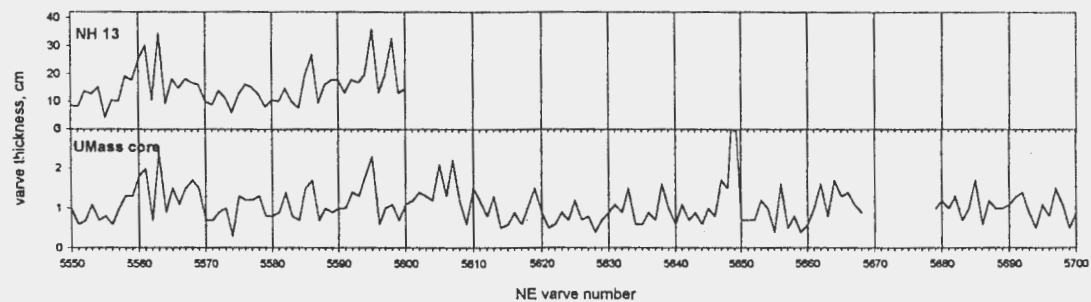
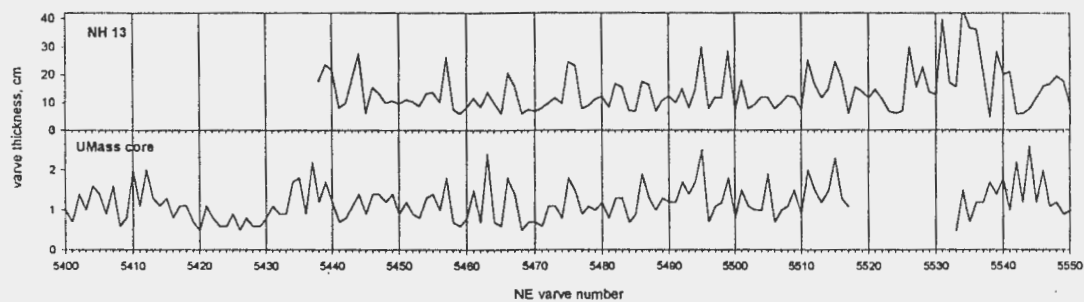
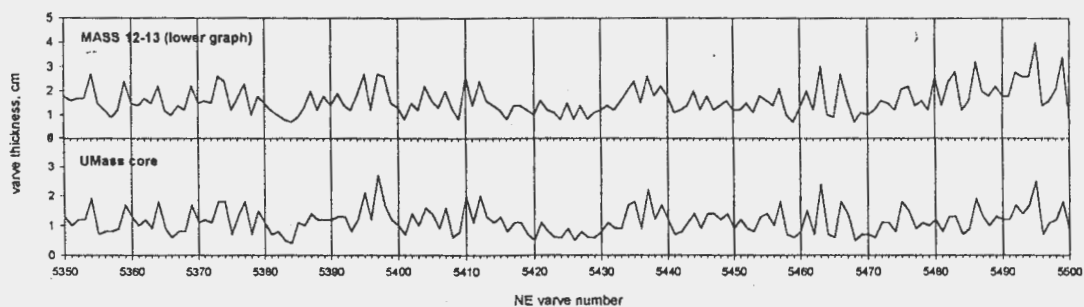
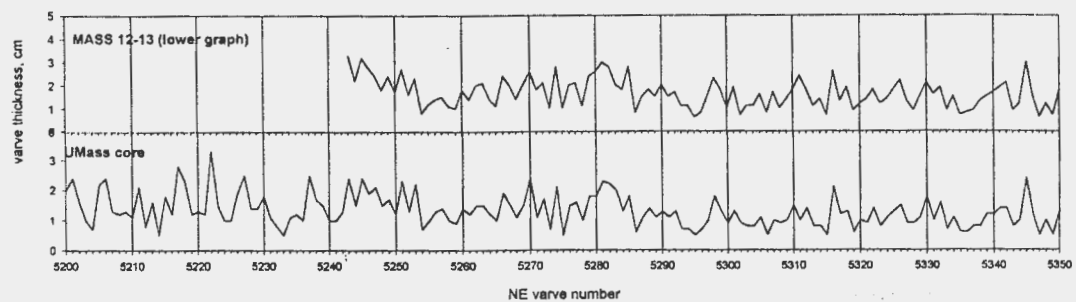
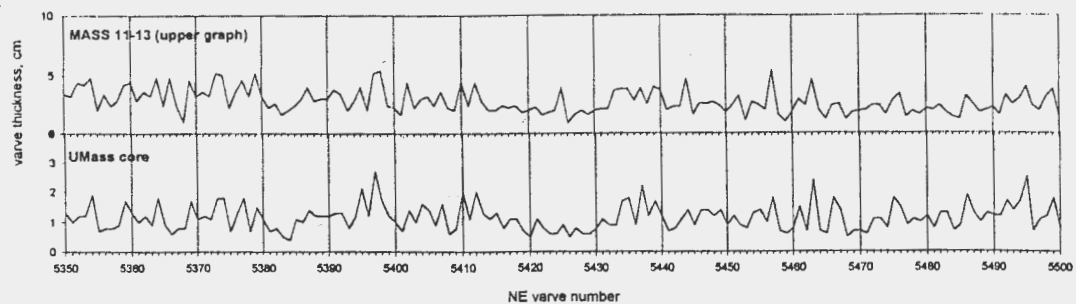
CORRELATION WITH THE NEW ENGLAND VARVE CHRONOLOGY

Varves from the UMass core were correlated to the New England (NE) varve chronology (Antevs, 1922, 1928) by matching varve couplet thickness plots with normal curves of the NE varve chronology (Figure 1-3B, next three pages). Features used to visually match the UMass core with the NE varve chronology include the sequence of peaks in varve thickness, relative heights of peaks, number of varves within peaks and the presence of abnormally thick varves. Varves from within the UMass core were correlated to normal curves from Glacial Lake Hitchcock (data sets MASS 4 – 13, NH 13, VT 14 – 15, and VT-NH 16 – 17; Antevs, 1922) and the Canoe Brook section of Ridge and Larsen (1990). A 300-year section of the UMass core was also matched with a normal curve from the Hudson River valley (data set NY 13 – 14; Antevs, 1922). Correlation between the UMass core and the NE varve chronology was good to very good with correlation coefficients of $r = 0.5 - 0.9$.

The UMass core covers NE varve 4638 – 6027 (Antevs arbitrarily began the NE varve chronology at varve 3001), a total of 1,389 varves. This is the longest continuous sequence of varves measured to date; it covers approximately 1/3 of the NE varve chronology. All other varve measurements have been obtained from natural exposures and are limited to the sides of the lake basin where streams have cut through the varved sediment.

Inconsistencies between the UMass core and the NE varve chronology were due to missing varves in liquefied portions of the core (29 varves), missing varves at core breaks (1 varve), erosion of varves in the lake due to slumps as indicated by zones of contorted sediments (37 varves), and local variations in sedimentation and erosion on the lake bottom (11 varves). NE varve 4645 – 4655 and NE varve 4664 – 4681 were missing or partially preserved near the base of the core because the sediments became liquefied during the coring process. NE varve 4759 was lost between core drives, although portions of other varves at core breaks may also have been lost. Contorted varve zones within the core were due to slumps on the bottom of the former lake. At the base of some of the contorted zones varves were eroded by the flowing sediment. For example it is clear NE varve 4705 – 4706, 4824 – 4826, 5518 – 5532, 5762 – 5767, 5817 – 5827, and 5885 – 5886 were eroded from the coring location during subaqueous slump events. When compared to the NE varve chronology, ten additional varves were missing from the UMass core. After closer inspection of the sediments, evidence for eroded varves, such as sand layers, or misinterpretation of thin winter clay layers, explained the discrepancy.





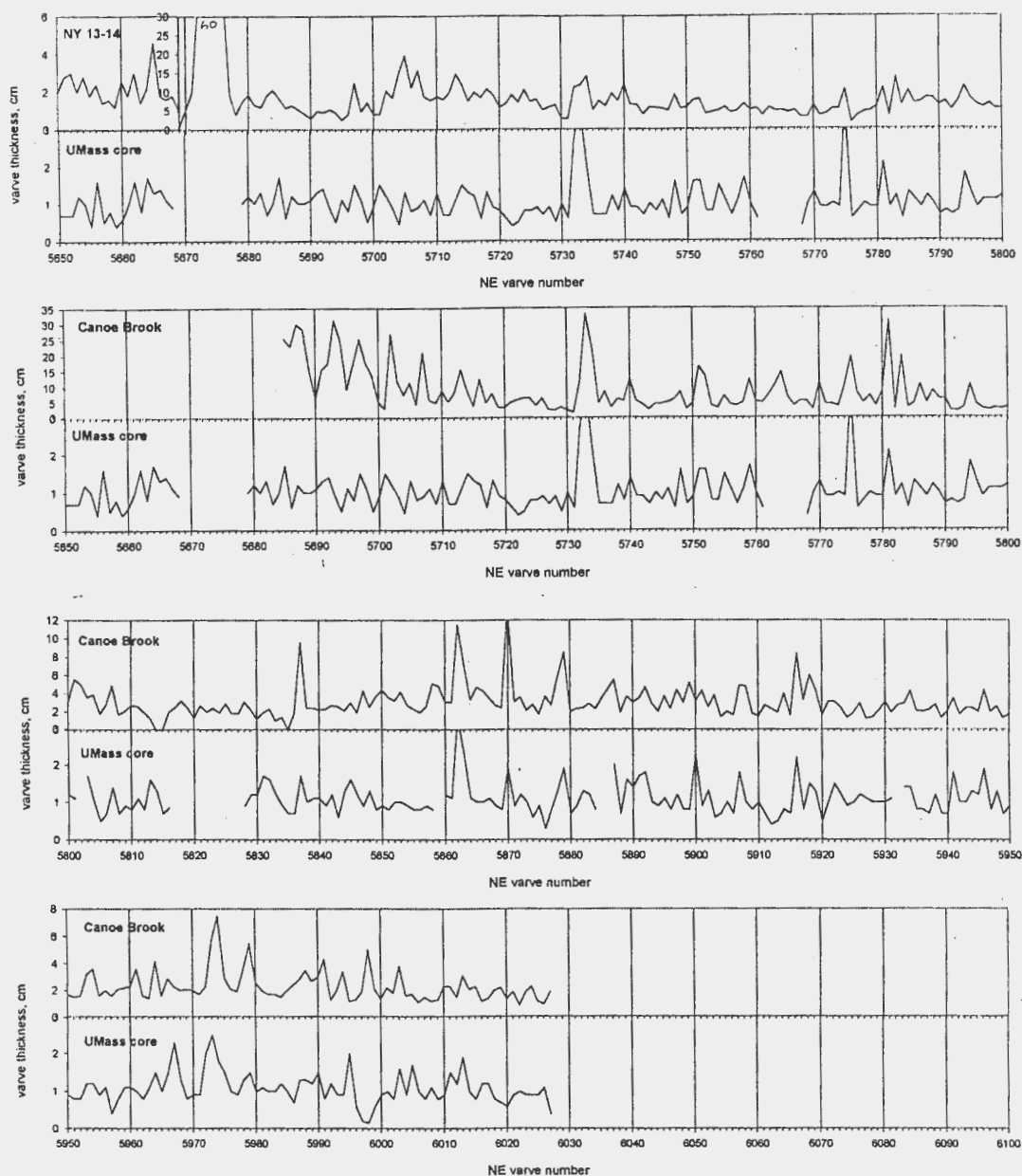


Figure 1-3B. (This and previous three pages) Varve couplet thickness plots from the UMass core correlated with normal curves of the NE varve chronology (Antevs 1922, 1928) and the Canoe Brook section (Ridge and Larsen, 1990). Varve thickness in centimeters verses NE varve number. Changes in thickness scale represented by additional Y-axis. Gaps in the UMass core are discussed in text.

The missing varves due to local sedimentological reasons are NE varve 4928, 5384, 5574, 5594, 5782, 5802, 5859, and 5932. One extra varve couplet not in the NE varve chronology was observed between NE varve 5514 and 5515. The lack of an outcrop-scale exposure of this varve interval hinders sedimentological interpretation of this extra varve couplet.

CORRECTION OF THE HUDSON VALLEY GAP

In 1922 when Ernst Antevs created the NE varve chronology, there were two gaps within the sequence. The first called the Claremont gap, was arbitrarily created by Antevs because he could not find a compelling correlation between varves in the lower (NE varve 3001 – 6352) and upper (NE varve 6601 – 7750) Connecticut River valley (Figure 1-3C). This gap has been corrected by Ridge *et al.* (1999) through

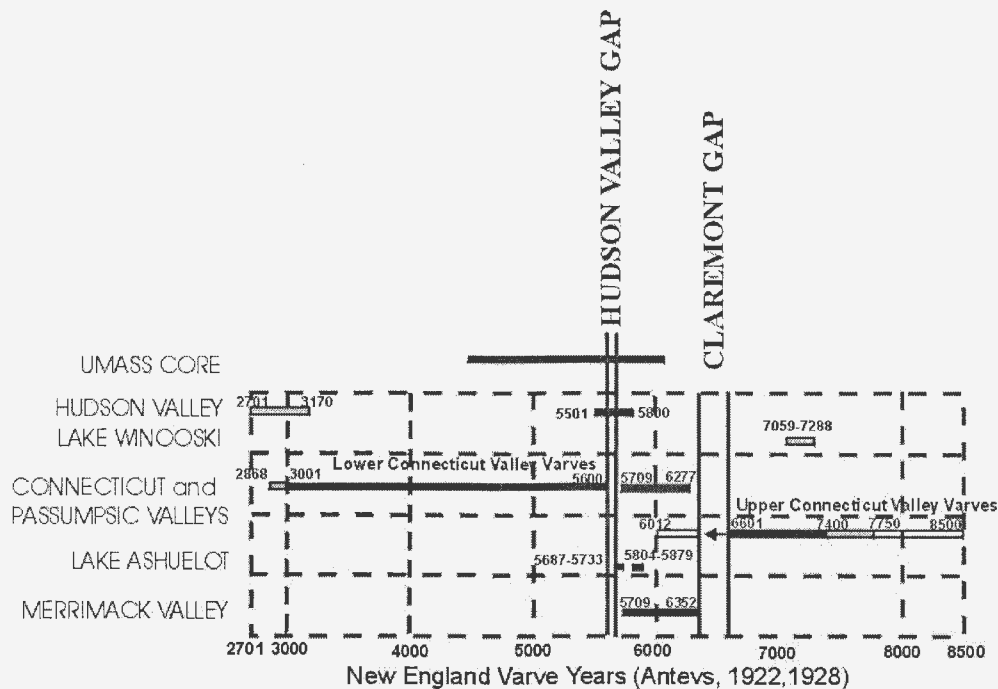


Figure 1-3C. Time span in varve years (younger to the right) of overlapping sequences of the NE varve chronology (NE 2701 – 8500) plus the UMass core (4638 – 6027). Hudson Valley gap (corrected here) and Claremont gap (corrected by Ridge and others, 1999) shown

measurements of additional varve sections and paleomagnetic investigations (lower NE varve 6012 = upper NE varve 6601). The second gap, known as the Hudson Valley gap,

was covered by the UMass core (Figure 1-3C). In this part of the NE varve chronology Antevs had to use varve sections measured by De Geer in the Hudson valley to span a gap (NE varves 5600 – 5687) in varves that could not be found in the Connecticut valley. Correlation between the UMass core and normal curve NY 13 – 14 has indicated that De Geer over counted his Hudson valley sequence at one point by 10 varves where flood events created a series of thick graded layers that he mistook for annual couplets (Figure 1-3D). Both the long and short core collected from the UMass campus covered this interval. Varve couplet thickness in the UMass cores matched the Hudson Valley record fairly well considering the records were from separate lake basins. However the UMass core was missing the 10 anomalously thick ‘varves’ counted by De Geer in the Hudson Valley. When these thick ‘varves’ are removed from the Hudson Valley record there is a good varve for varve correlation on either side of the flood event. Also there is no evidence for erosion or removal of varves in this position in the UMass core. Therefore,,

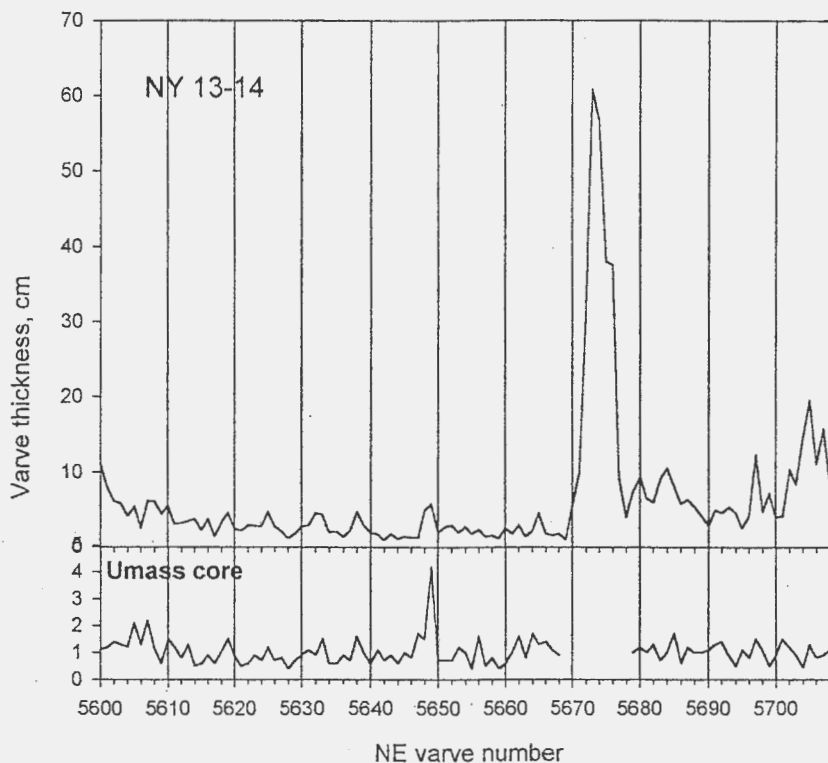


Figure 1-3D. Hudson Valley gap: correlation of the UMass core with NY 13-14 for the interval of the Hudson Valley gap (NE 5600 – 5709). The layered flood event that DeGeer mistook for annual couplets is the 10-point peak in the Hudson Valley record, NE varve 5669 – 5678. The UMass core correlates relatively well with the Hudson Valley varves on either side of the flood event and there is no evidence for eroded varves in the core at this interval. NE varves 5669 – 5678 have been removed from the NE varve chronology.

the 10 flood-event varves counted by De Geer, NE varve 5669 – 5678, have been removed from the NE varve chronology. These two small adjustments have been the only adjustments made to the NE varve chronology, the remaining varve thickness measurements and correlations in the over 4000-year varve chronology have withstood vigorous testing and have been shown to be correct.

VARVE SEDIMENTOLOGY IN THE UMASS CORE

Varves at the base of the core, are very thick (7 – 55 cm), red-tinted varves deposited directly upon mica-schist bedrock (Figure 1-3A). These basal varves were in an ice-proximal environment within Glacial Lake Hitchcock. Provenance studies of outwash plains, kame terraces, deltas and varves has shown that ice-proximal deposits within the Connecticut River valley are dominated by sediments derived from the local Triassic bedrock (O'Toole, 1988; Liu, 1989). In comparison, ice-distal deposits are primarily composed of crystalline lithologies from the surrounding uplands. These differences in provenance are due to changing forms of sediment delivery. In ice-proximal environments the predominant sediment source was from subglacial and ice-marginal glacial outwash. However, ice-distal environments were dominated by inwash from major tributaries entering the valley from the surrounding crystalline uplands (O'Toole, 1988; Liu, 1989).

As expected, the varve color changes from red, reflecting a local Triassic source in the thick ice-proximal varves, to grey reflecting a distal crystalline source in the thinner ice-distal varves (1 – 4 cm). Varve thickness gradually decreased throughout the UMass core from NE varve 4685 to 6027, except for an increase in varve thickness between NE varve 4879 – 4940 possibly due to the Camp Meeting Readvance (Emerson, 1898; Antevs, 1922). Varves from the upper part of the sequence, approximately NE varve 5700 – 6027, have more sand lenses in the summer layer and have thin, eroded winter clay layers. This suggests that these varves were deposited within a higher energy environment, possibly in the remnant lower-level Lake Hadley that existed in the basin during lake drainage.

At 1.5 m depth there is a gradational transition from thin (< 1 cm) poorly defined varves to non-varved lacustrine silt and sand (~ 20 cm thick). This lacustrine silt to medium sand unit commonly contains thin (1 – 2 mm) discontinuous clay layers. However these irregular silt and clay layers are not varves. Instead the discontinuous clay layers are interpreted as only remnant winter layers that were not eroded. These sediments were deposited during lake drainage when silt and sand was washed into the basin from rejuvenated streams and exposed shoreward deposits.

The uppermost portion of the UMass core (upper 1.2 m) is altered by soil development. The core location is near the present channel of Mill River. The upper 3.1 m of the longest core is composed of fluvial sand from this small stream. It is likely this

stream also eroded some of the non-varved lacustrine sand in the upper portion of the short core. The thickness of non-varved lacustrine sediment reported here is not likely to be the total thickness deposited during the existence of water within the Hadley basin.

The UMass core stratigraphy is typical of the varve stratigraphy seen throughout the lake basin: ice-proximal varves followed by ice-distal varves, with thin varves grading into non-varved lacustrine sediment at the top of the section. However, complete stratigraphic sections from bedrock/till up through non-varved lacustrine sediment are rarely exposed.

IMPLICATIONS OF THE UMASS CORE

Small plant fragments picked from varves in a contorted zone between NE varve 5761 – 5768 within the UMass core produced a $\delta^{13}\text{C}$ corrected AMS age estimate of $12,370 \pm 120$ ^{14}C yr BP ($^{13}\text{C}/^{12}\text{C}$ ratio: -27.1 ‰; Beta 124780). The $\delta^{13}\text{C}$ value (-27.1 ‰) for this sample is within the range expected for terrestrial plants and trees (-26 to -28 ‰) (Stuiver and Polach, 1977), suggesting the sample has not been contaminated by non-atmospheric ^{14}C from bedrock. When converted to calendar years this sample has an age estimate of $14.3^{+1.2}_{-0.4}$ cal kyr BP. This age estimate correlates well with the calendar year time scale placed on the NE varve chronology by Ridge *et al.* (1999) which suggests that NE varve 5768 was deposited 14.7 cal. kyr BP. This part of the varve chronology falls near a large ^{14}C plateau where 800 calendar years are represented by only 200 radiocarbon years. Due to the reduced production of ^{14}C between 15.2 and 14.4 cal. kyr BP and because the date only provides a maximum age for NE varve 5768, this radiocarbon age estimate provides only minimal temporal constraint on the NE varve chronology. However, its coherence with the NE varve time scale supports its validity.

At least 259 varves (NE varve 5768 – 6027) were deposited within the UMass core after deposition of the dated horizon, indicating that varve deposition continued within the Hadley basin until 14.0 cal. kyr BP. This suggests that Glacial Lake Hitchcock completely drained in the Hadley basin sometime after 14.0 cal. kyr BP.

Deglaciation of the Amherst area can be determined by projection of the dated horizon calibrated to calendar years to the base of the UMass core. This suggests that NE varve 4638 corresponds to 15.4 cal. kyr BP, indicating that the ice margin had retreated from the Amherst area and ice-proximal varves were deposited by this time. This age estimate of deglaciation is consistent with the Ridge *et al.* (1999) ice retreat chronology that places the ice margin near Amherst MA around 15.8 cal. kyr BP. Age estimates from the top and bottom of the UMass core indicates that it covers varve deposition from 15.4 to 14.0 cal. kyr BP.

References for Stop 3A.

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- Antevs, E., 1928, The last glaciation with special reference to the ice sheet in North America: American Geographical Society Research Series 17, 292 p.
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- Kuenen, P.H., 1951, Mechanics of varve formation and the action of turbidity currents: *Geologiska Föereningen i Stockholm Föerhandlingar*, v. 73, p. 69-84.
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- O'Toole, P.B., 1998, Provenance evidence for inwash and subglacial fluvial origins for the debris in Connecticut valley heads of outwash: Unpublished Masters Thesis, Department of Geological Sciences, Lehigh University, 171 p.
- Ridge, J.C., Besonen, M.R., Brochu, M., Brown, S.L., Callahan, J.W., Cook, G.J., Nicholson, R.S., and Toll, T.J., 1999, Varve, paleomagnetic, and ^{14}C chronologies for the late Pleistocene events in New Hampshire and Vermont, USA: *Geographie Physique et Quaternaire*, v. 53, p. 1-31.
- Ridge, J.C. and Larsen, F.D., 1990, Re-evaluation of Antevs' New England varve chronology and new radiocarbon dates of sediments from Glacial Lake Hitchcock: *Geological Society of America Bulletin*, v. 102, p. 889-899.
- Stuiver, M., and Polach, H.A., 1977, Discussion, Reporting of ^{14}C data: *Radiocarbon*, v.19, p. 355-363.

STOP 1-3B CLIMATE ANALYSIS RESULTS FROM THE NEW ENGLAND VARVE CHRONOLOGY References for this discussion are included in **Rittenour et al, 2000 at the beginning of the guidebook.**

Spectral analyses were conducted on the New England varve chronology (NE varve 2868 – 6900) to test for climate signals in the varve-thickness record. The results of these analyses indicate that there are prominent climate cycles in the varves (Article F, Figure 2). Statistically significant climate signals occur at multidecadal (>40 year) timescales, within the conventional 2.5-5 year cycles associated with El Niño/Southern Oscillation (ENSO), and at 7-9 year and the roughly 2.1-year "quasibiennial" period associated with the North Atlantic Oscillation, although the 2.1-year period signal should be interpreted cautiously because of its proximity to the 2-year Nyquist frequency (lowest resolution) for annual sampling. The 22-year period signal has the same frequency as the Hale solar magnetic cycle, which has been correlated with modern Northern Hemisphere temperature variations.

Examination of the climate signal through time with an evolutive spectral analysis, has indicated that the some of the climate signals change over time (Article F,

Figure 4). Most conspicuous is the change in the intensity of the ENSO signal (3–5 yr). In the first half of the record, ~17.5 to 15.5 cal kyr BP (14.6 to 12.9 ^{14}C kyr BP), the ENSO-like climate frequencies are prominent in the varve thickness record, while in the latter half of the record they are weak to diminished. A closer look at the record reveals that the high-frequency (2.5 to 2.8 yr) component of the ENSO-like signal is persistently strong throughout the New England varve chronology, while the lower frequency component (3.3–5 yr) is strong from 17.5 to 15.5 cal kyr BP and reduced from 15.5 to 13.5 cal kyr BP.

This long-term trend of continual weakening of the interannual variability throughout the 4000-year record appears systematic rather than random in nature and may be associated with low-frequency astronomical forcing of ENSO. The high-frequency (2.5–3 year) component of ENSO is typically associated with the lifetime of ENSO events and the El Niño/La Niña alteration, while the lower frequency component (3–7 year) is typically associated with the spacing of large events. The absence of the lower-frequency component of variability during the latter part of the interval studied here (i.e., the latest Pleistocene) is consistent with the theoretical prediction of fewer large events in the early Holocene. Our results suggest that the ENSO system was operational during the late Pleistocene, when the Laurentide ice sheet was near its maximum extent and climatic boundary conditions were different than today. The change in the low frequency component of the ENSO-band may suggest a weakening of the ENSO system after 15.5 cal kyr BP.

Despite the similarities between the ENSO pattern seen in the NE varve chronology and modern and paleoclimatic records (both have three spectral peaks in the 3–7 yr bandwidth), it is not possible to determine if the changing intensity of the ENSO-like signal is directly related to El Niño intensity in the tropical Pacific Ocean. One would expect ENSO to influence seasonal melt patterns in New England during the late Pleistocene. However, ENSO teleconnections into New England may have been continually altered as climate conditions changed during ice retreat. It is plausible that the retreat of the Laurentide ice margin, with its tendency to deflect storm tracks, could have led to a slow northward migration of storm tracks, and thus ENSO teleconnections into North America themselves, which represent a modulation of these storm track patterns. However, varve deposition in Glacial Lake Hitchcock followed the ice margin with progressively younger varves deposited northward, therefore limiting this effect. While the persistent high-frequency component of the ENSO-band throughout the New England varve chronology argues against any influence of nonstationary teleconnections, the possibility nonetheless remains that the observed changes are influenced by some combination of changes in the nature of ENSO teleconnections into North America and changes in ENSO itself. Our primary conclusion that ENSO was active during the late Pleistocene follows, however, in either scenario.

RoadLog. Continued: Exit Mullins Way and turn right onto Memorial Drive (west) and take Rte 116 North. Go north about 4 miles and turn right into the Warner Construction Gravel pit entrance. Do not enter the Gravel pit. (cumulative miles = 35.9 miles).

Stop 1-4 SUNDERLAND DELTA & ECHO DUNES – J. Brigham-Grette and Tammy Rittenour

The Sunderland Delta (or Long Plain Delta, Figure 1-4A) is a classic example of the ice- contact/meteoric delta in the valley graded to the stable level of Glacial Lake Hitchcock. The topset/foreset contact here has an elevation of 295 ft (89.9 m). Once the ice margin had retreated northward and bifurcated around Mt Toby to the north, melt waters and rains continued to flow into the lake across this delta.

Following drainage of Glacial Lake Hitchcock from the Hadley basin, the Long Plain Brook quickly adjusted to a lower base level by incising a small canyon through the delta (Figure 1-4B). This downcutting produced an alluvial fan out over the floor of the former lake. Today, groundwater maintains a continuous flow through the delta producing natural springs from sand and gravel beds confined by lake clay and silt at the foot of the delta slope. The State of Massachusetts maintains two fish hatcheries for trout at the base of the delta. These fish hatcheries are open to the general public during posted hours.

Periglacial conditions persisted in the valley for some time following lake drainage. The presence of ice wedge pseudomorphs, possible pingo scars and various dune forms found on the ancient lake floor and related landforms provides pervasive evidence of arid, cold conditions. Here in front of the Sunderland delta, winds blowing across the exposed sediments of Glacial Lake Hitchcock picked up sediments and created an echo dune (Clos-Arceuduc, 1969). Such dunes are formed due to air turbulence as wind rises over an obstacle, commonly a steep bedrock outcrop, drumlin, or delta front on the eastern side of the lake basin. These transverse dunes form strings of elongate dunes that conform to the planimetric shape of the obstacle. The distance of the echo dune from the obstacle is a function of the height of the escarpment (Quency and Dubief, 1943; Tsoar and others, 1983). Internal sedimentary structures include many stacks of small wispy blowout and deposition structures. These dunes have a nearly symmetrical morphology because wind eddies sculpted and deposited sediment on both the eastern and western sides of the dune (Figure 1-4C).

References for Stop 1-4

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- Quency, P., and Dubief, J., 1943, Action d'un obstacle ou d'un fosse sur un vent change de sable: Trav. Inst. De Rech. Sah, v.2, 169-176.
- Rittenour, T.M., 1999, Drainage history of Glacial Lake Hitchcock, Northeastern USA., Masters Thesis, Department of Geosciences, University of Massachusetts, Amherst, MA 178 pg.

Tsoar, H. Gurion, B., and Sheva, B., 1983, Wind tunnel modeling of echo and climbing dunes: In Brookfield, M.E., and Ahbrandt, T.S., eds., *Eolian Sediments and Processes*, v. 38, *Developments in Sedimentology*, Elsevier, New York, p. 247-259.



Figure 1-4A. Topographic map of the Sunderland Delta. Note narrow canyon cut by the Long Plain River as Lake Hitchcock drained and the strategic location of the fish hatcheries. Rte 63 cuts through the echo dune at the base of the delta.

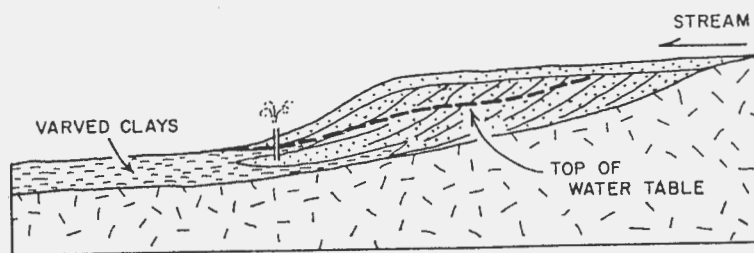
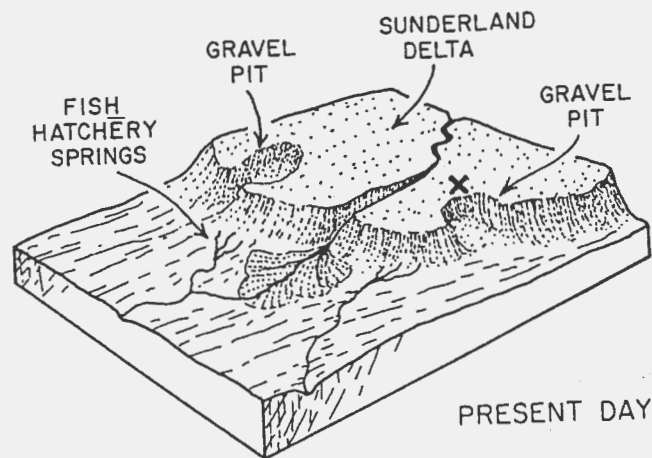


Figure 1-4B. Cartoon and schematic cross section of the Sunderland Delta showing the location of the alluvial fan at the mouth of the small canyon and the flow of ground water through the delta stratigraphy.

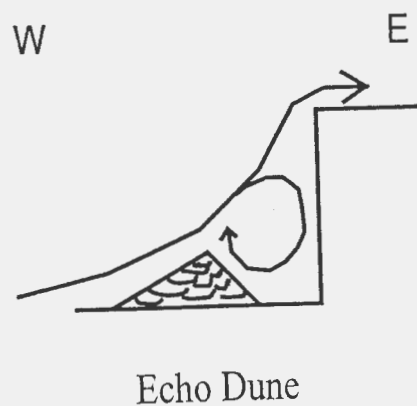


Figure 1-4C. Cartoon of echo dune morphology and formation.

Roadlog continues : Continue north on Rte 116 0.8 miles. The road here cuts through the ecdodune. continue north on rte 116 into Sunderland. Cross Rte 47 and the Connecticut river. Take immediate right onto River road. Continue north 1.8 miles and park on the right across from cattle ramp. (cumulative miles = 40.6 miles)

STOP 1-5 RIVER ROAD VARVE SITE--A. Werner, L.B. Levy and J.

Brigham-Grette

Site: The UMass River Road Varve site located just north of Sunderland (Figure 1-5A) lies on the west side of the Connecticut River and on the eastern flank of the Pocumtuck Range (Mount Sugarloaf). The site lies midway between the Montague Plain (topset/foreset contact = 102.7 m) and Long Plain delta (89.9 m). The inferred lake level at this site is 95.1 m (see attached map). Based on this water level the water depth at this site would have been ca. 50 meters and the shoreline would have been approximately 1

km to the west. Two small slumps along this small creek expose over 200 years of sedimentation into glacial Lake Hitchcock.

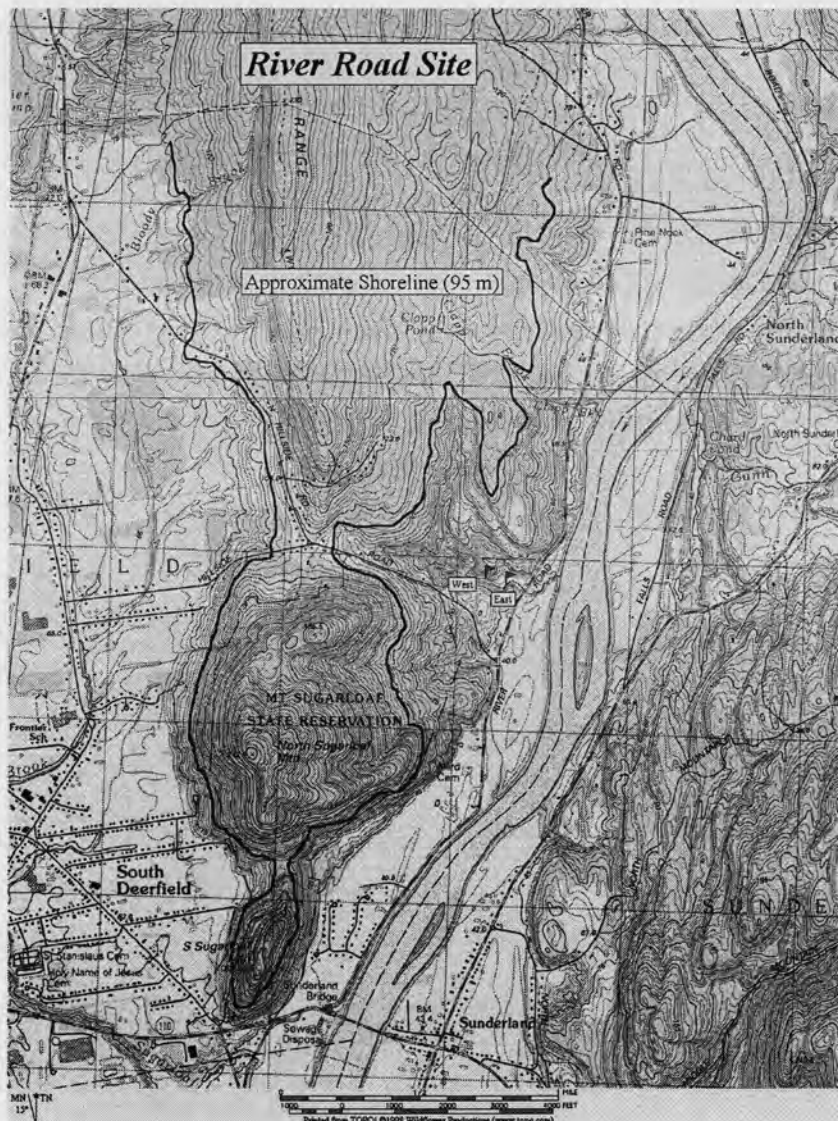


Figure 1-5A. Location of the varve section on River Road on the east side of the Pocumtuck Range in South Deerfield. The land here is owned by UMass.

Varves: The average varve thickness is 3 cm but varve thickness varies from <1 cm to over 10 cm. The two

exposures have been correlated to each other (Levy, 1998) and to the Antev's master curve (varve #s 4980 to 5200) using patterns of varve thickness (Thomas, 1984). Absolute thickness values do not correlate well, however, the curve patterns show a remarkable similarity. Using Jack Ridge's calibrated varve curve the age range for the site is inferred to be ca. 13,988 - 13,146 yr B.P. There is a positive (but weak) correlation ($r^2 = .29$) between summer silt and winter clay thickness and strong correlation between summer silt and total varve thickness ($r^2 = .96$), demonstrating the relative thickness of the silt layers (Figure 1-5B). The varve stratigraphy shows an overall thinning up-section, perhaps, indicating increasing distance to the glacier margin. Sand layers occur periodically in the exposures, most often capping the winter clay layers (Spring deposition) and the summer silt layers (Fall deposition). Disturbed beds (7 - 40 cm thick), occurring as a homogenous mix of silt and clay or as contorted varve layers, are observed at both sites – they are more numerous at the western site but are thicker at the Eastern site. These beds are interpreted as subaqueous slumps off the highland to the west due to increased runoff, wave agitation or paleoseismicity.

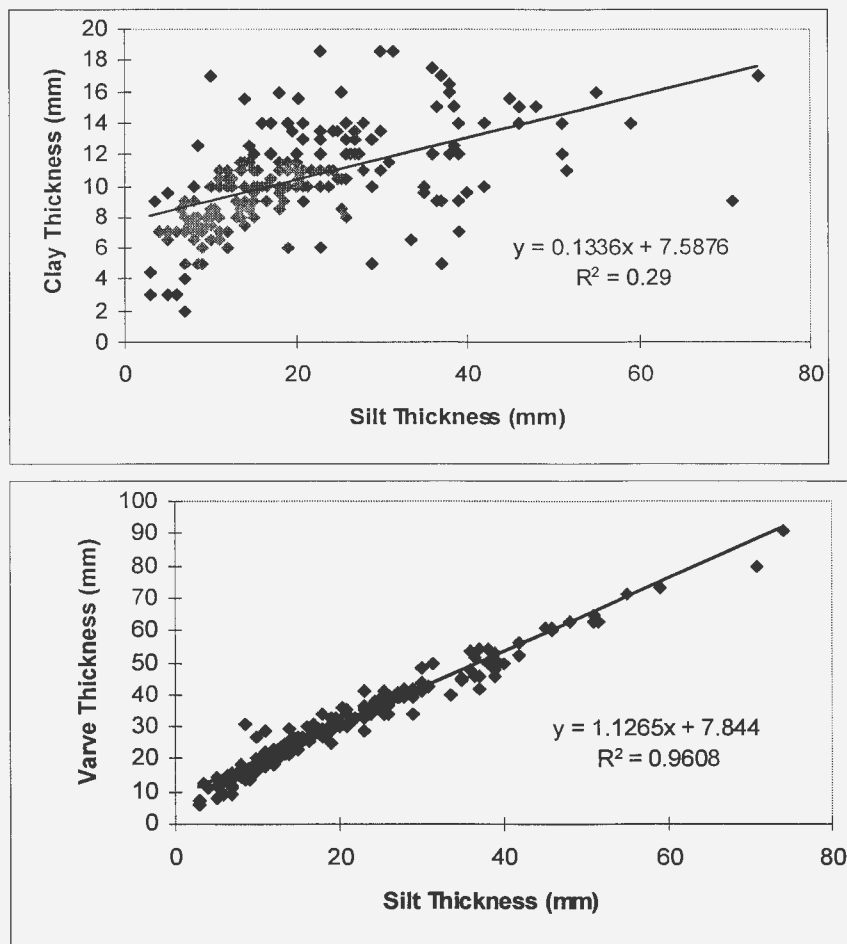


Figure 1-5B. The upper diagram demonstrates the high correlation between changes in the “summer” silt layer and total varve thickness. The lower diagram indicates a low correlation between the thickness of the “summer” silt layer and the “winter” clay layer.

Concretions: Carbonate (40-45 wt. %) concretions occur in lake and marine sediments throughout the Northeast (Pardi, 1983). At the River Road Site, they are found within 5 distinct intervals in the western exposure, and in a single layer in the eastern exposure. Concretion layers at the western exposure do not occur in the eastern exposure and the only concretion zone in the eastern exposure does not correlate to a layer of concretions at the western exposure. The concretions found at the River Road exposures range in size from 1 to 5 cm (long axis) and their morphologies appear to be closely related to the microstratigraphy of the host sediment. Concretions in homogeneous disturbed zones have an ovoid shape, concretions found within well-layered varves have morphologies that reflect the texture of the enclosing sediment and concretions found in contorted disturbed beds mimic the micro structure of the enclosing sediment. Host sediment laminae do not change thickness through a concretion suggesting that the carbonate cement passively filled-in the pore spaces subsequent to dewatering and compaction.

Most of the concretions found at this site appear to be weathering as evidenced by an outer “rusty” rind, and most have a remnant carbonate core (although concretions high in the section are almost completely weathered). In polished section the concretions show original host sediment laminae and concretionary banding – no nuclei have been observed. The unweathered boundaries between the concretionary bands appear as thin (0.5mm) white zones that oxidize readily to form rusty dendritic boundaries. These boundaries are associated with elevated manganese levels (SEM-EDS data) presumably reflecting changes in pore water chemistry during concretion growth. The source of the concretion carbonate is thought to be carbon originating from terrestrial organic matter deposited with the lake sediment based on carbon isotope data. Individual concretion bands have been AMS ^{14}C age dated to evaluate their utility to provide reliable age control. Concretion centers date up to 2,500 years too old and the outer concretion samples too young (relative to the inferred age of the site) questioning the reliability of this approach. Recent work by Bosiljka Glumac (Smith College) indicates oxygen isotope ratios of -8.5 to -13.0 ‰ (SMOW) - values consistent with interglacial (local) meteoric water.

(some room for notes_____)

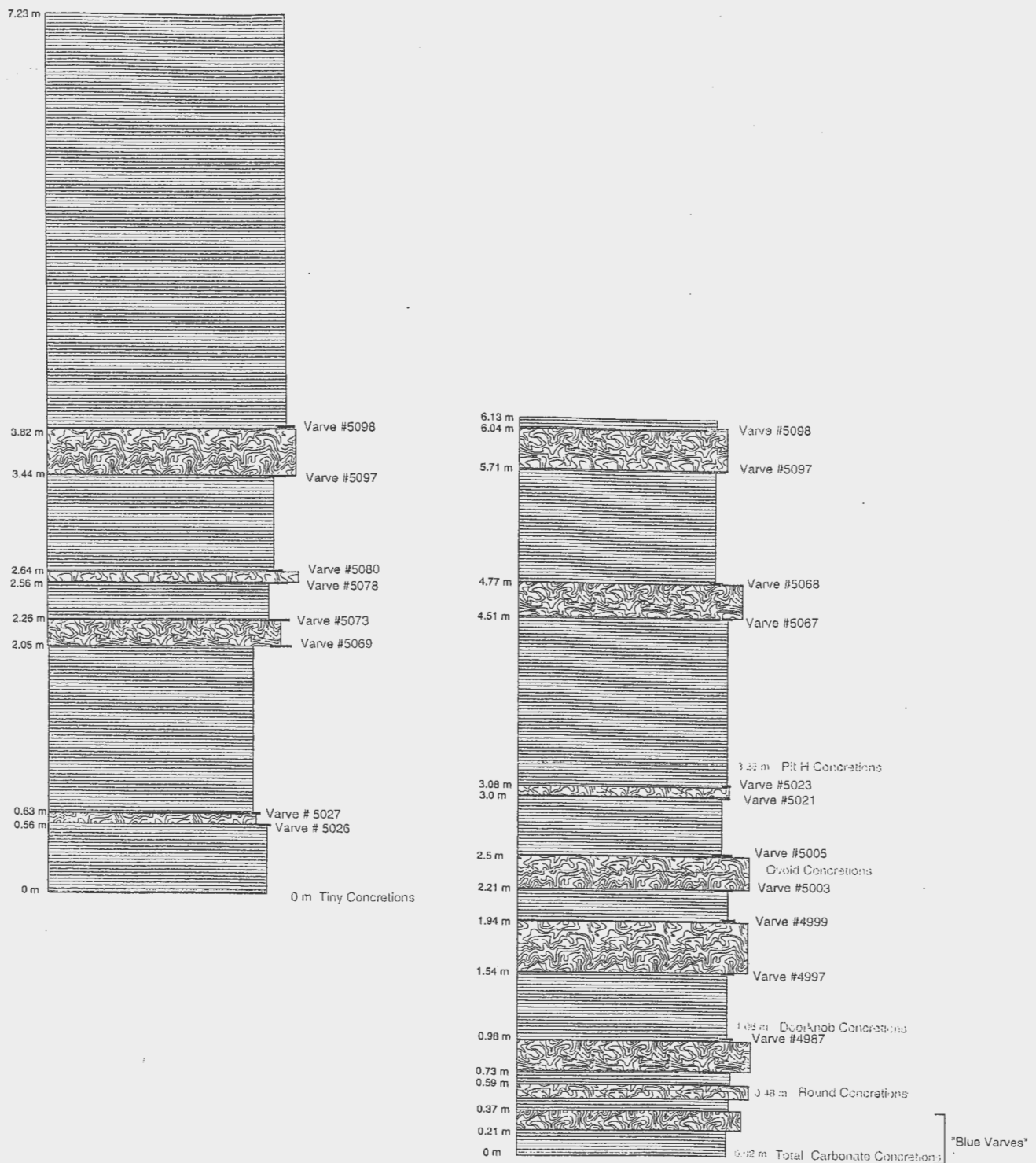


Figure 1-5C. Stratigraphic columns of the River Road East and River Road West exposures. (Levy, 1998, pg. 24-25).

References:

- Levy, L.B. (1998), Interpreting the carbonate concretions of Glacial Lake Hitchcock. Unpublished undergraduate thesis, Mount Holyoke College, South Hadley Massachusetts.
- Pardi, R.R., (1983), The origin of carbonate concretions within glaciolacustrine and glaciomarine sediments of the Northeast United States. Unpublished Ph.D. dissertation, University of Pennsylvania, Philadelphia, Pennsylvania.
- Thomas, G.M., (1984), A comparison of the paleomagnetic character of some varves and tills from the Connecticut Valley. Unpublished M.S. thesis, University of Massachusetts, Amherst, Massachusetts.

Road Log Continues: Go back south to Rte 116. Take a left and cross bridge, continue to light at the intersection with Rte 47. Take a left onto Rte 47 and go north 4.7 miles to intersection with Rte 63. Turn left onto Rte 63 and continue north 2.5 miles and take left onto Lake Pleasant Road. You are now on top of the Montague Delta. Take this road 2.3 miles to its end where it intersects Millers Falls Road. Take left onto Millers Falls Road and note sand dunes on the delta surface at the Turners Falls Airport (OSL dates in this area noted at Stop 1-2). Follow Millers Falls road to Main Street in downtown Turner's Falls. Take a right on Main Street and cross the bridge over the Connecticut River to the intersection with Rte 2 at the light. Take a right followed by an immediate right onto the perimeter Road of Riverside village. (cumulative miles = 56.6 miles)

Stop 1-6 GEOCHRONOLOGY FROM ARCHAEOLOGY: ALLUVIAL TERRACES IN THE RIVERSIDE DISTRICT OF GILL, MASSACHUSETTS by Kathryn Curran,

ABSTRACT: Archaeological sites on successive glacial and alluvial surfaces provide a chronology for the formation of postglacial Connecticut River channels and terraces. After Glacial Lake Hitchcock drained, the river carved a channel through Montague Delta deposits. Near Turner's Falls, Massachusetts, the Connecticut swung north and west of a bedrock ridge, combining with the Falls River. Archaeological evidence from river-cut terraces indicates that this initial channel was abandoned by 10,000 BP, as the Connecticut River established its present course. Geomorphic processes are accurately timed with archaeological data.

Archaeological sites on successive glacial and alluvial surfaces provide a chronology for the formation of postglacial Connecticut River channels and terraces just above Turner's Falls, in the Riverside District of Gill, Massachusetts (Figure 1-6A). The current course of the river and western jog between Northfield and Greenfield is a recent

feature that did not exist prior to the last glaciation (Brigham-Grette and Wise: 1988: 234).

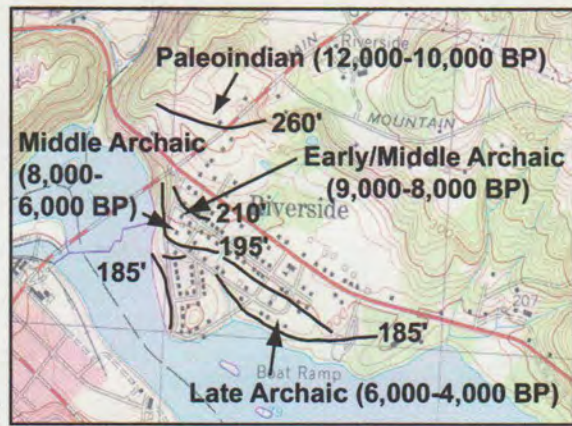


Figure 1-6A Riverside Terraces

Between 16,000-17,000 BP, the Laurentide ice sheet began to lying Connecticut River Valley. By 15,000-16,000 BP the stagnating

ice margin reached central and northern Massachusetts (Brigham-Grette and Wise 1988: 210). In front of the ice, Glacial Lake Hitchcock continued to expand northward. The present-day Riverside District was below the surface of the lake, estimated to be approximately 350-ft ASL (Figure 1-6B) as based on the elevation of a complex of deltas in the region (Brigham-Grette and Wise 1988: At some point in the life span of the lake, the sediment dam at Rocky Hill, Connecticut eroded and the pooled meltwater began to drain. Presently geologists are debating the definitive drainage chronology. The section at issue here, in northern. Massachusetts, drained by 13,000 BP (Ridge and Larsen 1990: 898). Between 14,000 and 12,000 BP, the Connecticut River began to erode sediments deposited in its former channel during glaciation and deglaciation (Jahns 1947: 29).

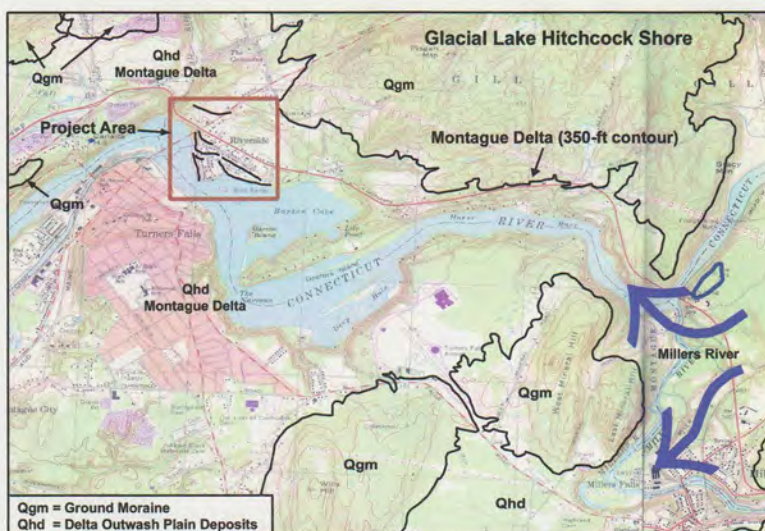


Figure 1-6B. Extent of Hitchcock at 350 ft elevation line.

The postglacial drainage channel flowing towards Miller's River encountered the Montague Plain Delta (Figure 1-6C). Thick, coarse delta deposits and bedrock prevented the river from reoccupying its earlier direct south-trending channel. The river diverted around the

As downcutting of the proto Connecticut continued, river level fell beneath 200-ft ASL and water no longer crossed over the Lily Pond Barrier (Figure 1-6E). The river followed a course similar to that seen today, west over Turner's Falls, remaining east of Canada Hill and then flowing southward. The associated terrace between 210-ft and 195-ft in elevation has two important archaeological sites. A C-14 sample collected from a site found between 200-ft and 210-ft yielded a date of 8685 ± 370 years BP (Peter Thomas 1997: personal communication). The terrace was inhabited between the end of the Early Archaic (10,000-8,000 BP) and the beginning of the Middle Archaic (8,000-6,000 BP).

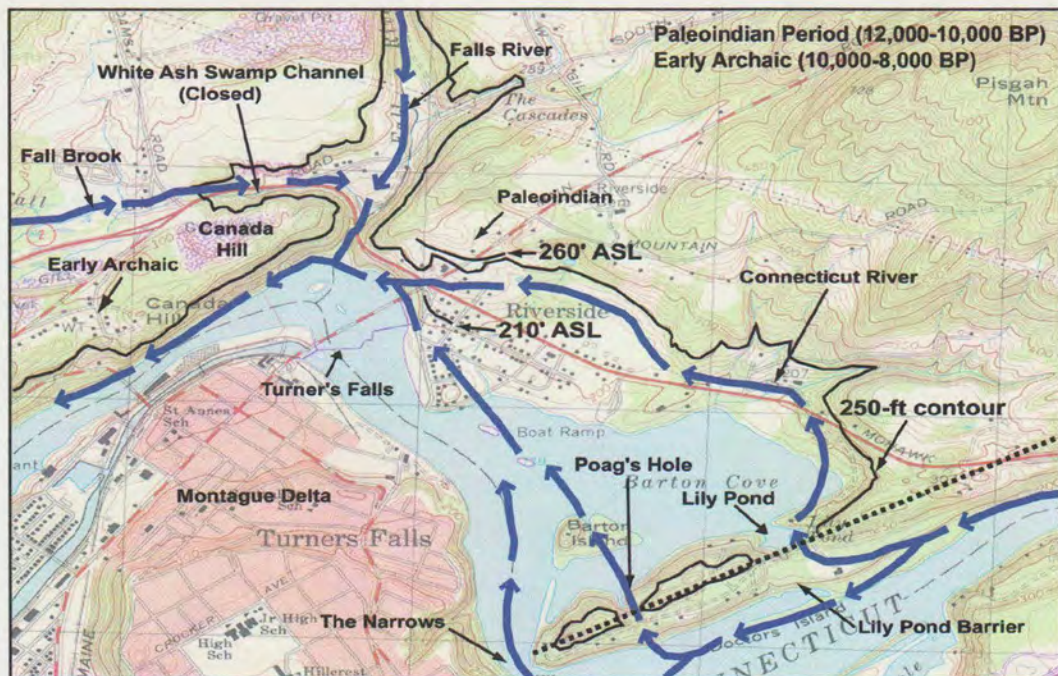


Figure 1-6D The White Ash Swamp channel is closed. Paleoindian and Early Archaic cultural materials are associated.

A second site is located between 195-ft and 200-ft in elevation (Figure 1-6E). Middle Archaic (8,000-6,000 BP) projectile points including Stark (7,000 BP) and Neville (7,800 BP) varieties were recovered. On the basis of information from these two archaeological sites, I speculate that human populations occupied this terrace as early as 9,000 BP. Water level dropped and a steep scarp near 195-ft was carved sometime after 7,800 BP. Since the Lily Pond Barrier was abandoned at 200-ft in elevation, I suggest that the plunge pools were active from the 11th millennium BP to approximately 8,000 BP. Archaeological evidence from Riverside terraces allows a decent chronology of postglacial downcutting by the Connecticut River to be established. Radiocarbon and known cultural chronologies date the use of these terraces by native Americans, and thus

Road Log continues: Return on road to Rte 2 Go right and continue 1 mile. Turn right into Barton's Cover Recreation Area, operated by Northwest Utilities. (cumulative miles = 58.0 miles)

STOP 1-7A BARTONS COVE ;Gill Massachusetts, south of Hwy 2, SE of Riverside, MA; 702000m E, 4719500m N UTM; Greenfield Quadrangle; Led by Tammy Rittenour

Sequence of Lake Drainage

During the Stable phase of Lake Hitchcock, the Montague Delta (fed by the Millers River) extended completely across the lake basin. This is evident from the delta foreset directions along the west side of Canada Hill and north of the Mineral Hills (Figure 1-7A-1). The Montague Delta separated the narrow, shallow lake basin to the north (in the Northfield MA area) from the deep, broad Hadley basin to the south. In turn the Hadley basin was separated from the remaining portions of Glacial Lake Hitchcock to the south by the large Chicopee-Westfield Delta complex, which also grew to extend across the lake basin.

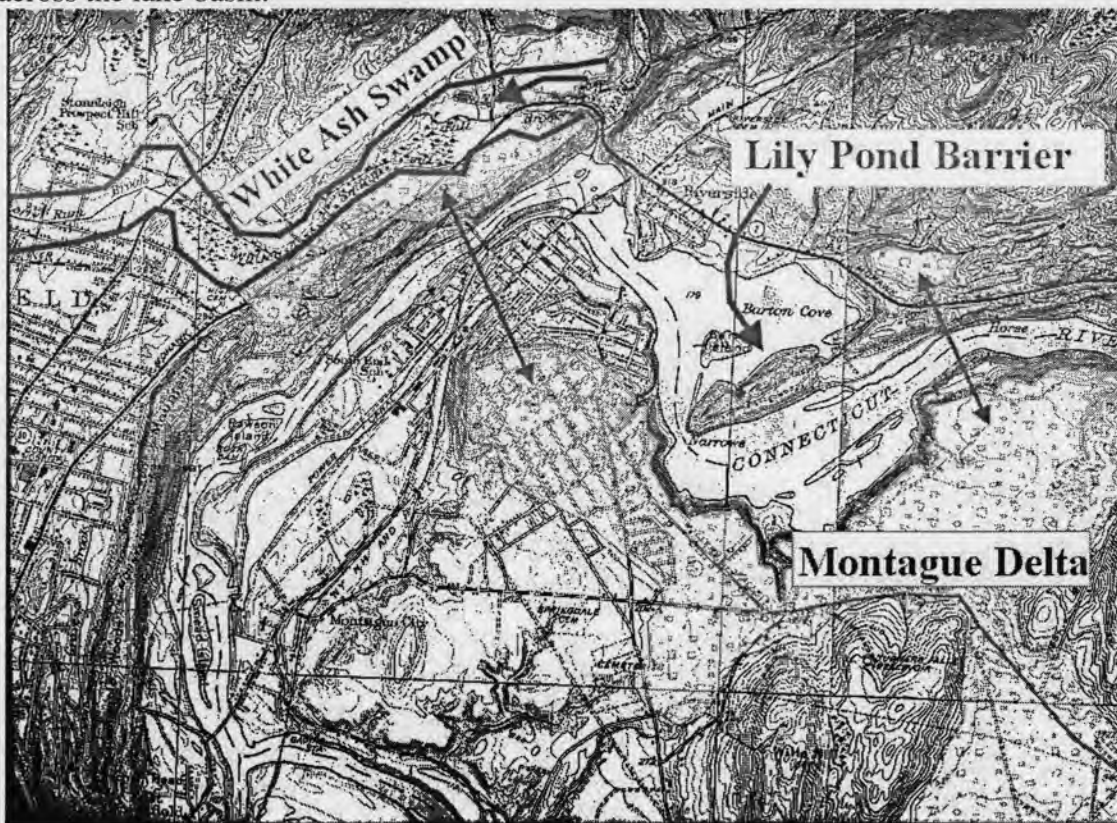


Figure 1-7A-1. Topographic map of the Montague Delta area, Greenfield Quadrangle, 10 foot contour interval. Patterned area represents the once continuous surface of the Montague Delta. The White Ash Swamp channel and Lily Pond Barrier are labeled

Lake levels north of the Mount Holyoke Range were controlled by incision into the Rocky Hill dam and subsequent entrenchment of the proto-Connecticut River into the Chicopee-Westfield Delta complex. The first lake level drop in the Hadley basin is recorded in topset/foreset contacts in the southernmost stretch of the Montague Delta that are 3-5m below the Stable lake level (Jahns and Willard, 1942; Koteff and Larsen, 1989). This drop in lake level may or may not correlate with the terrace cut into the Rocky Hill dam at 2-5 meters below the Stable lake level and the other deltas south of the Mount Holyoke Range built at this level (Koteff and others, 1988; Koteff and Larsen, 1989; Stone and Ashley, 1992).

During initial lake drainage a proto-Connecticut River channel was developed on the west side of Canada Hill (Figure 1-7A-1). This channel, called the White Ash Swamp channel, has a channel base of 84 m asl. (270 ft asl) (~20 m below the stable lake level here) and is the oldest surface occupied by the Connecticut River during incision into the Montague delta. The White Ash Swamp channel was abandoned during further river incision and development of a more favorable route, similar to that of the modern Connecticut River, on the east side of the Canada Hill bedrock ridge.

LAKE HADLEY

Terraces cut into the Rocky Hill dam at 12 and 9 m (8 – 11 m below the stable level of Glacial Lake Hitchcock) represent complete drainage of the lake south of the Chicopee-Westfield Delta (Stone and Ashley, 1992). After progressive headward erosion into the Chicopee-Westfield Delta complex, a lake 20 m below the stable level, known as Lake Hadley (Jahns, 1967) developed in the deep Hadley basin (Figure 1-7A-2). Evidence for the existence of Lake Hadley has come from a remnant delta deposited into the northern extent of the lake basin in the vicinity of Sunderland MA (Jahns, 1967), and subtle wave-cut benches near North Amherst and South Hadley, MA. In addition, well defined terraces 20 m below the stable lake level are found cut into the Chicopee and Westfield deltas by the Connecticut, Chicopee and Westfield Rivers.

During the existence of Lake Hadley, abundant silt and sand was washed into the relatively shallow lake basin (~ 20 m deep) from rejuvenated streams and exposed shoreward deposits. This lacustrine sand unit is seen throughout the former extent of Lake Hadley and can be as much as 12 m thick. Aprons of in-washed silt and sand are found near Easthampton and South Hadley, MA. The apex of most of these fans is near the former water level of Lake Hadley, locally around 60 m asl (200 ft asl).

Stratigraphic descriptions of lake bottom sediments in the Hadley basin indicate a transition from thick grey Lake Hitchcock varves to thinner varves, then lacustrine sand and silt deposition on top (Jahns and Lattman, 1962; Jahns, 1967). In the UMass core the transition from upper varves with thin clay layers to lacustrine silt and

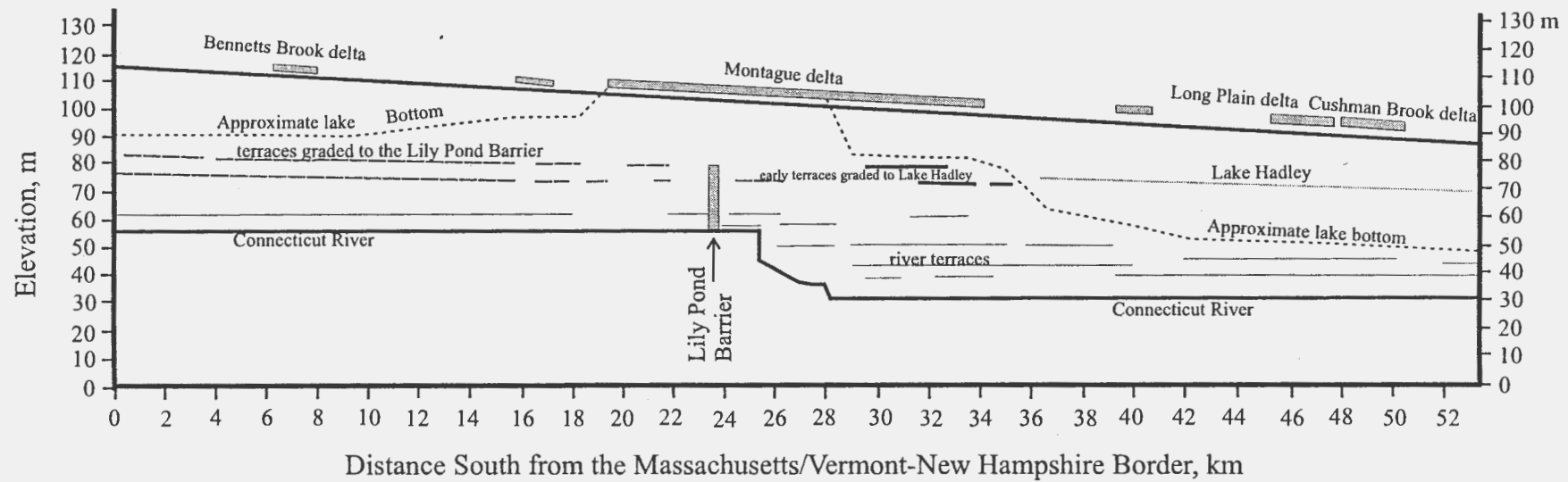


Figure 1-7A-2. Longitudinal profile of the Hadley basin, in Lake Hitchcock

sand deposition is gradational. However, Jahns (1967) described a section near Hatfield, MA, where the base of the lacustrine silt and sand is composed of contorted and ripped-up varve clasts and rippled sand indicating a current during deposition. Jahns (1967) interpreted laterally continuous fine-grained layers in the lake bottom sand as winter layers and under this assumption counted 30 ± 10 years worth of deposition in 6 m of sediment. Judging from estimates of unit thickness, Jahns (1967) suggested Lake Hadley could not have existed for much longer than 50 years.

TERRACES GRADED TO LAKE HADLEY

The Montague delta once extended across the lake basin separating the Hadley basin from the lake to the north. During the existence of Lake Hadley (20 m below the stable level) the shallow lake basin north of the Montague delta in Massachusetts and southern most Vermont/New Hampshire had drained. The proto-Connecticut River began to flow between a residual lower-level lake upstream in VT-NH and Lake Hadley. Early terraces were cut into the exposed lake bed and Montague delta sediments to meet the local base level - Lake Hadley. These early terraces, graded to Lake Hadley, are further evidence for the existence of a lower lake in the Hadley basin.

During incision of the Montague delta by the proto-Connecticut River, the Deerfield River entrenched northward into its large delta to meet the level of the Connecticut River. Remnant terraces throughout the Deerfield valley indicate the Deerfield River quickly planed off its delta surface to the level of the short-lived early strath terraces graded to Lake Hadley. Pre-formed channels on the delta surface may have caused the curious path of the Deerfield River to the north, instead of to the south where the base level of Lake Hadley could have been quickly met.

DRAINAGE OF LAKE HADLEY AND DEVELOPMENT OF THE LILLY POND BARRIER

Complete entrenchment through the Chicopee-Westfield delta complex eventually drained the Hadley basin. At this point early strath terraces cut by the flow of water across the lake bottom were formed. Within a short period of time, channelized flow was initiated in the proto-Connecticut River, and these early subtle terraces were abandoned. The exposed lake bottom sand quickly became subject to eolian transport, generating dunes and sand sheets. To the north, the drop in base level from the drainage of Lake Hadley caused the Connecticut River to further entrench into the Montague delta. During incision the river became perched on a bedrock ledge near Turners Falls MA (Figure 1-7A-2). Erosion continued south of this nick point but was suspended to the north. This situation created two waterfalls over the sandstone ridge with associated plunge pools at the base (Figure 1-7A-3 and Figure 1-7A-4). Prior to the flooding of Barton's Cove by the Turners Falls

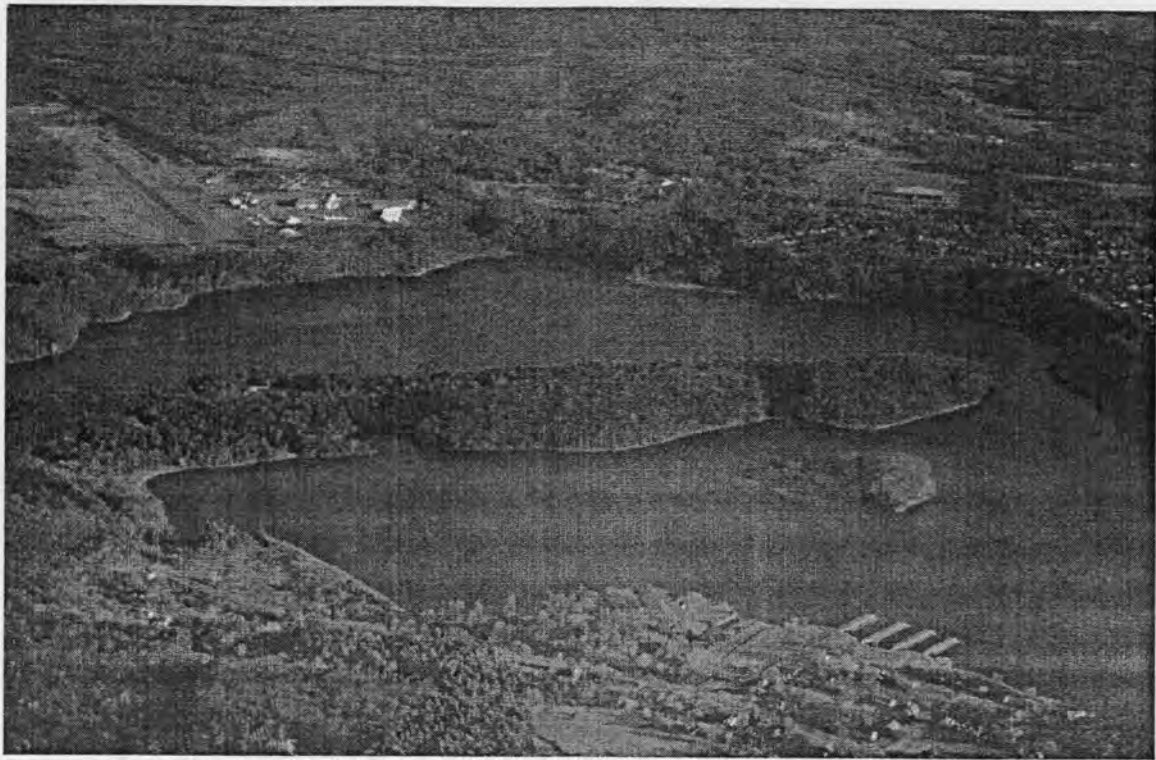
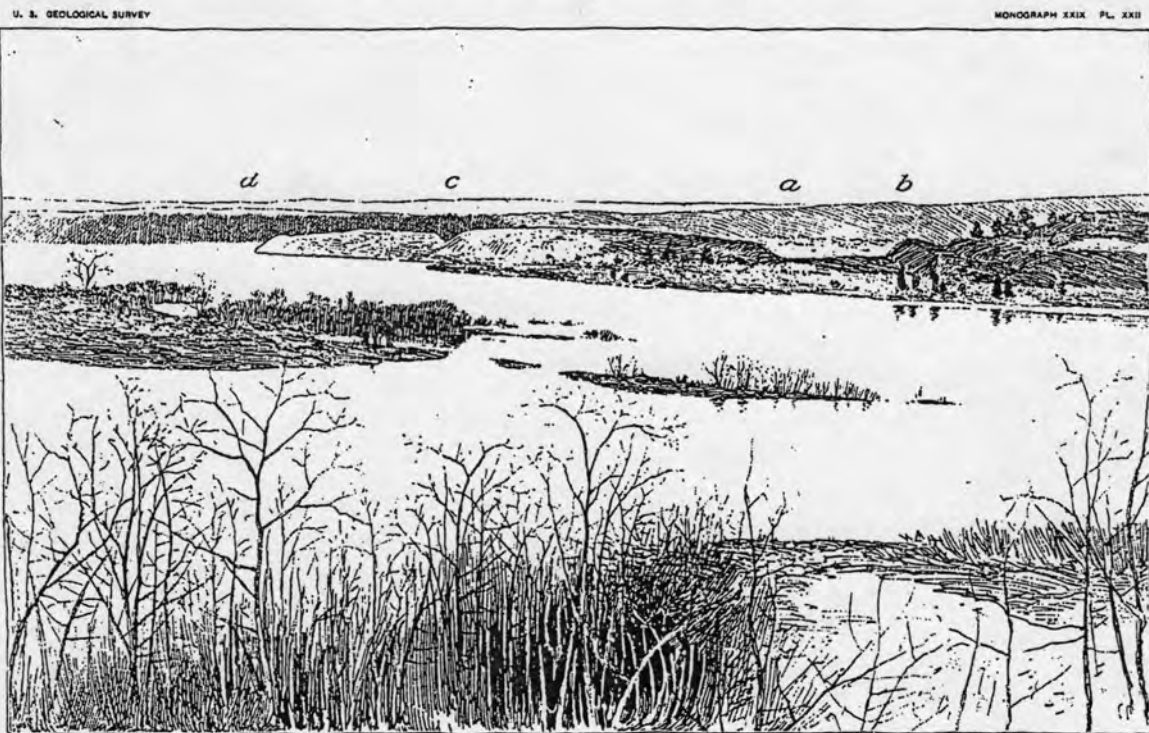


Figure 1-7A-3. Areal Photograph of the Lily Pond Barrier looking southeast.



VIEW ACROSS THE CONNECTICUT RIVER, SHOWING THE NOTCHES FORMED BY THE RIVER AT THE LILY POND, IN GILL, AND ITS ESCAPE AROUND THE NARROWS.
The water passed over the notches in the sandstone ridge below *a*-*b* and *c* as a waterfall directed from the observer, and later cut around the ridge at *d*.

Figure 1-7A-4. Drawing of Lily Pond Barrier in the early 1800s when the ride was deforested.

dam, these plunge pools were known as Lily Ponds and so the bedrock ridge has become known as the Lily Pond Barrier (Emerson, 1898). Deep channels were cut into the barrier and prominent terraces formed upstream as the Connecticut River flowed over this nick point (Figure 1-7A-5). Geomorphic and archeological evidence suggest that this nick point may have been active for 1000's of years, possibly having been abandoned some time before 10,200 cal. yr BP (9,000 ^{14}C yr BP) (Curran, 1999).

While the Connecticut River was perched on the Lily Pond Barrier, local base level to the north was the height of this nick point. Many of the residual lakes in the northern reach of the Glacial Lake Hitchcock basin at 8 – 10 m, 20 m and 30-40 m below the stable level could not drain until after the Connecticut River had eroded around the bedrock ridge and the nick point was abandoned. Varve deposition in a residual lower-level lake near Newbury, VT, ceased at 12,300 cal. yr BP (10,400 ^{14}C yr BP) (Ridge and others, 1999). This final lake drainage may have been due to abandonment of the Lily Pond Barrier and incision into the lake bottom in VT-NH.



Figure 1-7A-5. River terraces in the Northfield MA area that are graded to the Lily Pond Barrier

References Stop 7

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- Ridge, J.C., Besonen, M.R., Brochu, M., Brown, S.L., Callahan, J.W., Cook, G.J., Nicholson, R.S., and Toll, T.J., 1999, Varve, paleomagnetic, and ¹⁴C chronologies for Late Pleistocene events in New Hampshire and Vermont, U.S.A.: Geographie physique et Quaternaire, v. 53, p.1-31.
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Road Log continues: continue east 3.9 miles on Rte 2, cross over the French King Bridge and then turn left onto Rte 63. Go north on Rte 63 2.1 miles and turn left into the Northwest Utilities Picnic Area. (Cumulative miles = 64.0)

STOP 1-7B ABYSSAL DEPTHS IN TURNER'S FALLS AREA, FRENCH KING BRIDGE ; Ed Klekowski

How deep is the Connecticut River? Would it surprise you to know that in some places its depth exceeds 125 feet? It surprised me. Two deep sites in Massachusetts are currently being explored by divers. The following map shows their locations; both are near where the French King Bridge spans the Connecticut River. Site B, known as French King Hole, has been dived for years by Valley Divers. The second site, labeled Site A, known as King Philip's Abyss, was recently (summer, 1997) discovered and explored by University of Massachusetts divers. Both sites should be dived with caution: they are very deep, totally without light (dark as the inside of a cow!), have overhangs, and there is the possibility of entanglement from waterlogged, uprooted trees (roots, trunk and

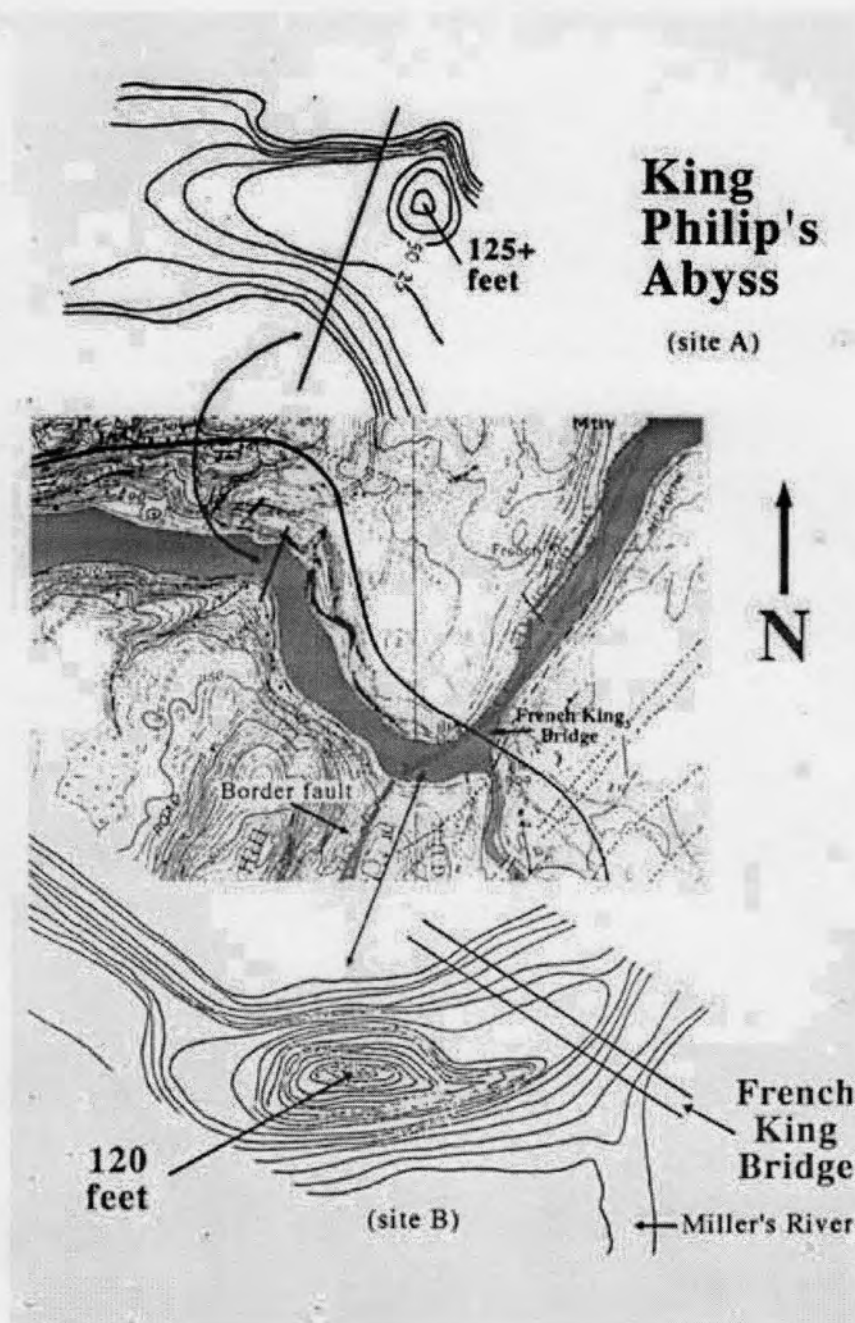


Figure 1-7B-1. Location of the French King Bridge across the Connecticut River and the deep eroded cliffs found below river level . One of the giant potholes coincides with the Eastern Border fault.

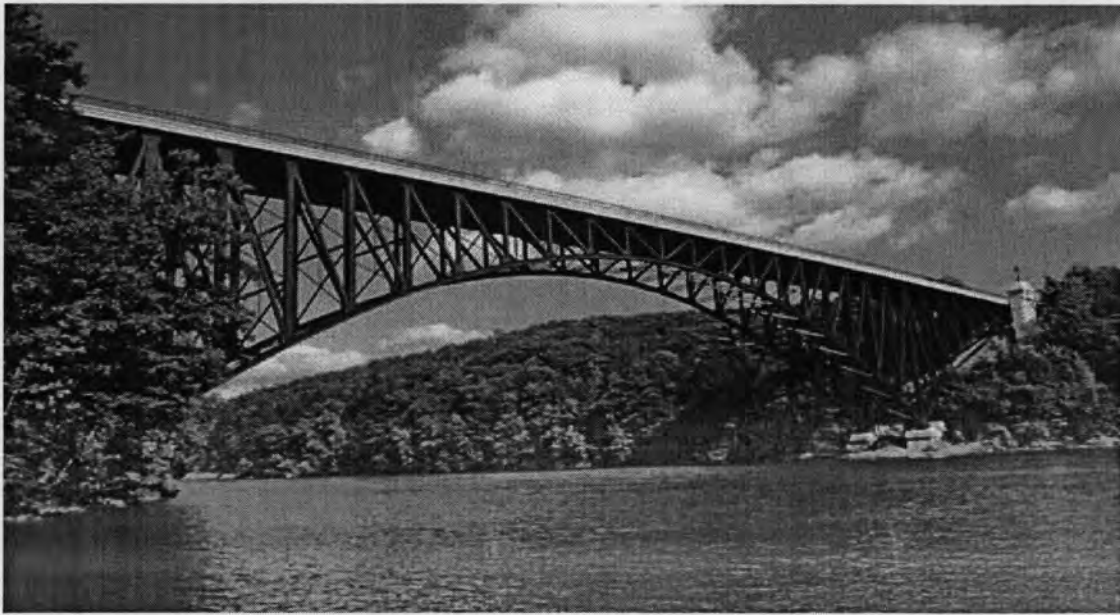


Figure 1-7B-2. THE BORDER FAULT CROSSES THE RIVER IN THE FOREGROUND UPSTREAM FROM THE FRENCH KING BRIDGE. THE BORDER FAULT IS ON THE RIGHT (EASTERLY) SIDE OF THE RIVER

crown) that sink to the bottom. As you will note on the map, both of these deep sites are close to a fault that crosses the Connecticut River. This fault, known as the Border Fault, is worth a geological digression.

Approximately 200 million years ago, all of the continents were united into a single supercontinent called Pangea. Subsequently faults separated the continental plates and they began the slow drift to their present locations. During the initial phase of this continental break up, a fault (the Border Fault) developed through what is now the Connecticut River Valley. This fault aborted, and another occurred east of Boston separating North America from Africa. Had the Border Fault not aborted, Boston would be in Africa and Greenfield, Massachusetts would be a coastal resort! Too Bad.

French King Hole (Site B) exactly coincides with where the Border Fault crosses the river. Faulting may result in neighboring rocks either becoming stronger or weaker. According to Professor Little, a geologist at Greenfield Community College, the rocks in the French King Gorge have been weakened by the Border Fault. Perhaps the 125' depth of this sector of the river is due to the erosion of these fault-weakened rocks by great volumes of water that have flowed through the Gorge in the geological past.

King Philip's Abyss is near the Border Fault, but no known fault crosses the river at this location. The origin of the Abyss is geologically unresolved, but its proximity to the Border Fault is tantalizing.

Of the two deep (125 feet plus) river sites, King Philip's Abyss is the more interesting biologically. Therefore our discussion will focus on this habitat. The Abyss is an underwater cliff that crosses the river. The edge of the cliff is 20-30 feet beneath the surface and plunges almost straight down for 80 feet in some locations. The bottom of the cliff is abrupt and then begins an unexplored slope of gravel and boulders going deeper. In some places the cliff is undercut into cavern-like areas.

Since the cliff is in the lee of the flow of the river, this habitat is unique. The cliff wall is festooned with life: white sponges as big as dinner plates and colonies of the bryozoan, *Lophopodella carteri*, a state listed species, cover the cliff face between the sponges.

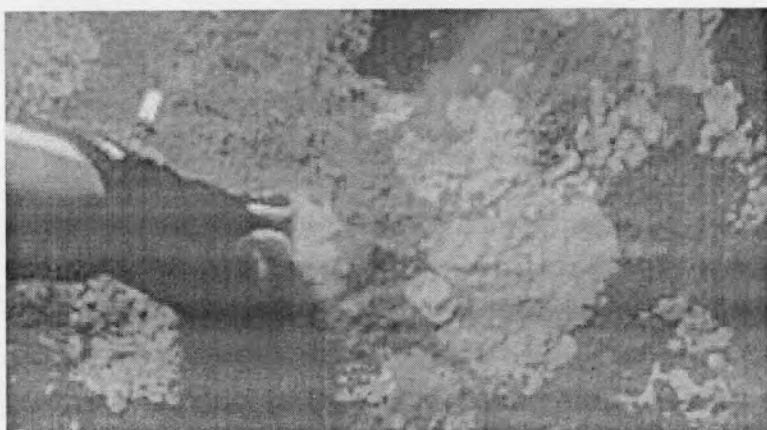


FIGURE 1-7B-3. SPONGES GROWING ON ABYSSAL WALL.
NOTE DIVER'S ARM FOR SCALE.

We as yet know very little about this deep water cliff-face community of invertebrates. During any given dive, because of the depths, only 10-15 minutes are spent sampling for organisms; thus the total diving time actually spent studying this unique environment has probably been less than one hour. We really don't know what is living down there! This might be the last environment in Massachusetts whose biota is unknown.

References for Stop 1-7B.

- Hamblin, W. Kenneth. 1985. The Earth's Dynamic Systems, 4th Edition. Macmillan Publishing Company, New York. See Chapter 22 for a clear discussion of continental drift.
- Stopen, Lynne E. 1988. Geometry and deformation history of mylonitic rocks and silicified zones along the Mesozoic Connecticut Valley border fault, western Massachusetts. Master's Thesis, Department of Geology and Geography, University of Massachusetts.

Road Log continues: Return to Rte. 63, take a left (north) and continue for 5.1 miles to intersection with Rte 10. Take a left onto Rte 10 and continue 1.2 miles (cross the Connecticut River) and stop of the side of the road. (Cumulative miles = 70.4 miles).

STOP 1-8 NORTHFIELD TERRACES AND KING PHILIP'S HILL, NORTHFIELD -- Dena F. Dincauze

The Post-Hitchcock sequence of river terraces in Northfield contrasts strongly with the situation in Riverside, Gill, downstream from the French King Bridge. In Northfield the sequence generally is two high-level terraces (locally called Plains) just below the Hitchcock lake beach, a drop of $60 \pm$ feet, then two lower-level terraces locally called Meadows; the lower is the modern floodplain. Archaeological sites on the Plains are mostly older than 5000 ^{14}C yrs BP, that is, the entire early Holocene set of sites occurs at the high elevation. The Meadows typically have sites of the last 3000 years. The 60' bluffs that separate the Plains from Meadows were excavated between 6000 and 3000 ^{14}C yrs BP.

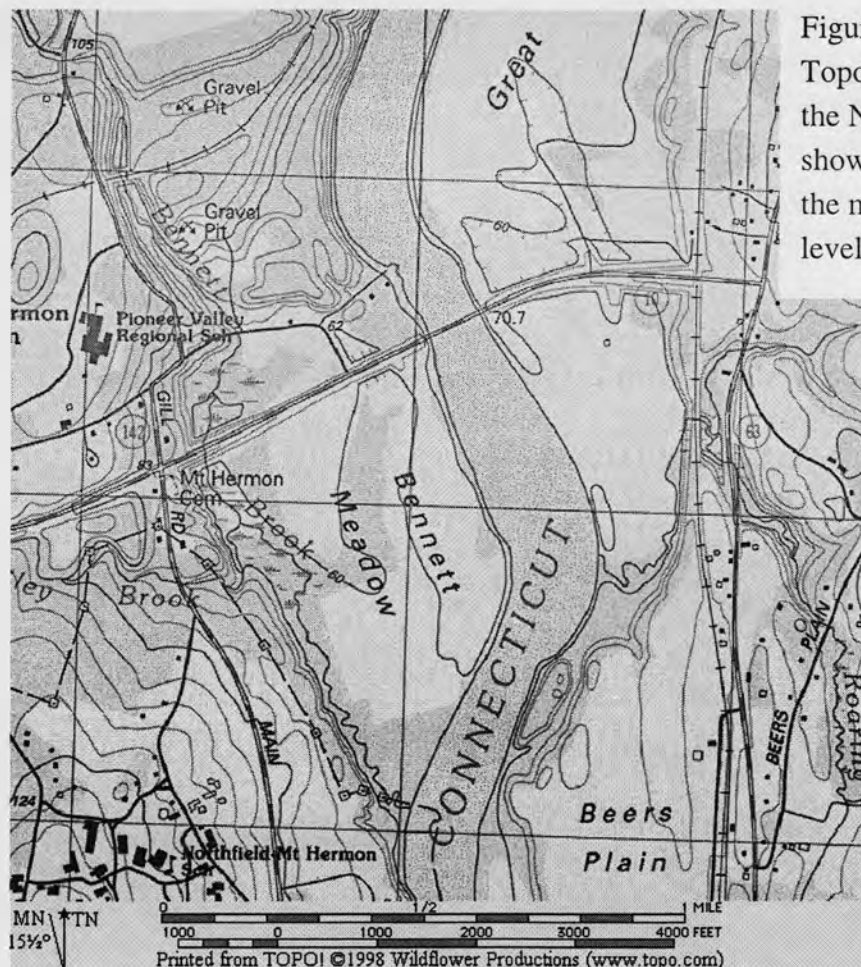


Figure. 1-8.A.
Topomap of part of
the Northfield Basin
showing several of
the major terrace
levels.

A spur of land above the western abutment of the former Bennett Meadow Bridge, north of the Rt. 10 Bridge, preserves small bits of both high Plains, with an alluvial fan on the higher one (K. Campbell's surficial map of Northfield). Across the spur is a narrow, straight trench with a low embankment on the river (east) side, which was given the name of "King Philip's Hill" in Northfield's rich legends. The story was that here King Philip built a fortified camp after his encounter with the Mohawks. The UMass archaeological field school in 1976 cut an exploratory trench at right angles through the ditch and bank to test our less legendary hypothesis that the feature was a seventeenth-century field boundary. Indeed, the ditch had a flat bottom and sharp corners, formed by metal shovels, and the bank was originally less than twice its present height. These specifics conform to the standard seventeenth-century field boundaries that occur elsewhere in the Connecticut Valley and eastern Massachusetts. This feature is one of the last tangible remains of the Second Settlement of Northfield.

As the archaeologists dug westward from the ditch they encountered, within intact sediments, two prehistoric artifacts: a rough biface and a small hammerstone. These items are not closely diagnostic of a particular culture, but both are in historic context only in the early Holocene-- between 9000 and 7000 years ago. They were incorporated into a fluvial deposit which included clay balls transported from the bed of Glacial Lake Hitchcock, and which cut into the gravel of an older alluvial fan, forming a low cliff west of the ditch. The artifacts were originally deposited on a surface intersected by the highest river channel in this stretch, but of Early Holocene age. The implication seems clear--something downstream was holding up the river on the surface of the lake beds, for a very long time. The Lily Pond Barrier? The French King Gorge?

Road Log Continues: Continue 3.5 miles west on Rte 10 to I-91 (Interstate Exit 28). Take I-91 South to Exit 18 and back to the Inn at Northampton. (Cumulative miles = 102.4 miles)

DAY 2: FIELD TRIP STOPS AND ROAD LOG

References for Stops 2-1 to 2-4 are listed in Ridge and others (1999) at the beginning of the guidebook

From Northampton, MA head north about 40 miles on Rt. 91 toward Vermont to EXIT 4: Rt. 5 in Putney, VT. Turn left at the end of the exit ramp to Rt. 5. The mileage log begins at Rt. 5.

Total miles (increment)

- 0.0 (0.0) Turn left (south) on Rt. 5.
1.7 (1.7) After entering valley of Canoe Brook pull over to right side of Rt. 5 at entrance to small saw mill. Do not block driveway. Cross Rt. 5 on foot (BE CAREFUL!!!) to small dirt access road leading to clay pit on north side of Canoe Brook.

STOP 2-1: Canoe Brook Varve Section, Dummerston, Vt. NOTE: Admittance to this exposure is by permission only from Sweet Tree Farm, Inc. in Vernon, Vt. and requires the signing of a liability release form.

The Canoe Brook section has been the best exposure of varves in southern Vt and N.H. for the last 15 years. The clay and silt excavated here is mixed with granular slate fragments to produce a topping for clay tennis courts. The 645-yr varve sequence was measured (Ridge and Larsen, 1990) and the bottom 550 varves match the New England (NE) Varve Chronology of Ernst Antevs (1922) at NE varves 5685-6234. This part of Antevs' chronology was compiled from varve outcrops in the Hudson, Merrimack, and Ashuelot (Keene, NH) valleys in addition to exposures in the Connecticut Valley from Putney to Charlestown (Figures 2-1 A and B). The upper 95 couplets are sandy and have load deformation features that prevent valid measurements for the purpose of correlation.

The base of the exposure has thick (30 cm) sandy couplets with well laminated summer layers that appear to represent a time shortly after deglaciation (within 50 yr). In addition to sandy sediment and high sedimentation rates an ice-proximal environment is supported by couplets containing very faintly laminated to massive clayey silt beds that appear to have settled from dense plumes in the presence of weak bottom currents. These units are interbedded with well-laminated sandy beds that are sometimes rippled and were deposited in the presence of strong bottom currents. In addition the basal 40 couplets lacks ostracodes which are easily separated from remaining parts of the section. Varve thickness decreases up section and drops to less than 2 cm where couplets 400 yr from the base are dominated by winter clay beds. Other features of the stratigraphy include abnormally thick varves that contain giant (10 cm) graded beds of clayey silt (NE 5733, 33.2 cm; NE5862, 11.5 cm). These units occur basin-wide and appear to represent catastrophic discharge events produced by the release of water from ice-dammed lakes in tributaries up valley. The varves at Canoe Brook were deposited in water depths of 58-42 m if they were deposited in the stable phase of Lake Hitchcock (Koteff and Larsen, 1989). Water depth was at least 30 m if the waning stages of varve deposition occurred in the Cold River stage (10-12 m below stable phase, see Stop 2-3).

In addition to varve stratigraphy the Canoe Bk. section provided the first calibration of Antevs' (1922, 1928) New England Varve Chronology by yielding the first ^{14}C ages (Table 2-1) from a matched varve sequence. Previously published ^{14}C ages are from near the top of the

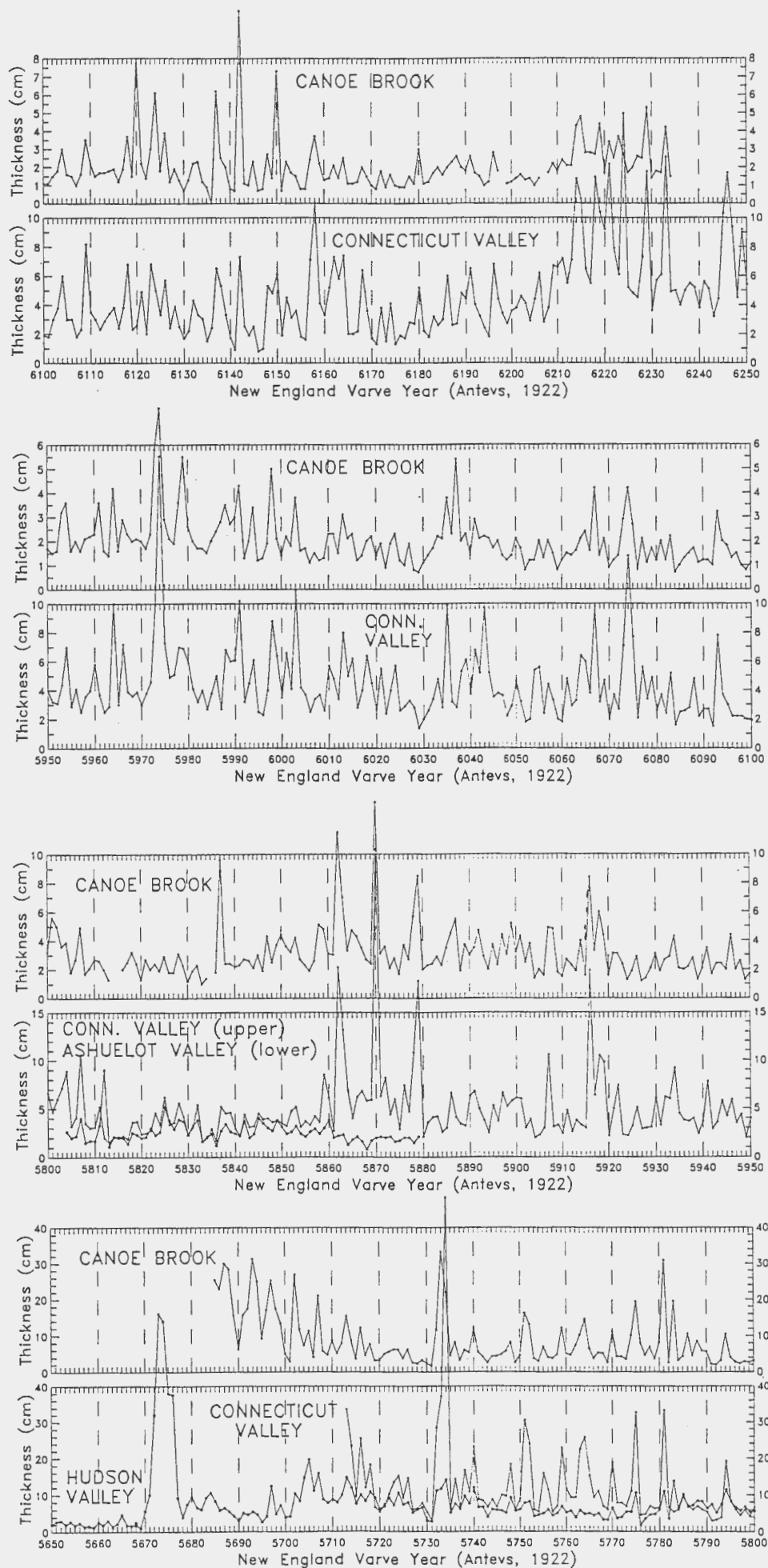


Figure 2-1A. Comparison of varve record at Canoe Brook (top; Ridge and Larsen, 1990) with matching sequences in the Connecticut Valley, Ashuelot Valley at Keene, NH, and Hudson Valley (bottom; Antevs, 1922). Graphs are 150-yr sections of New England varve years 5650-6250.

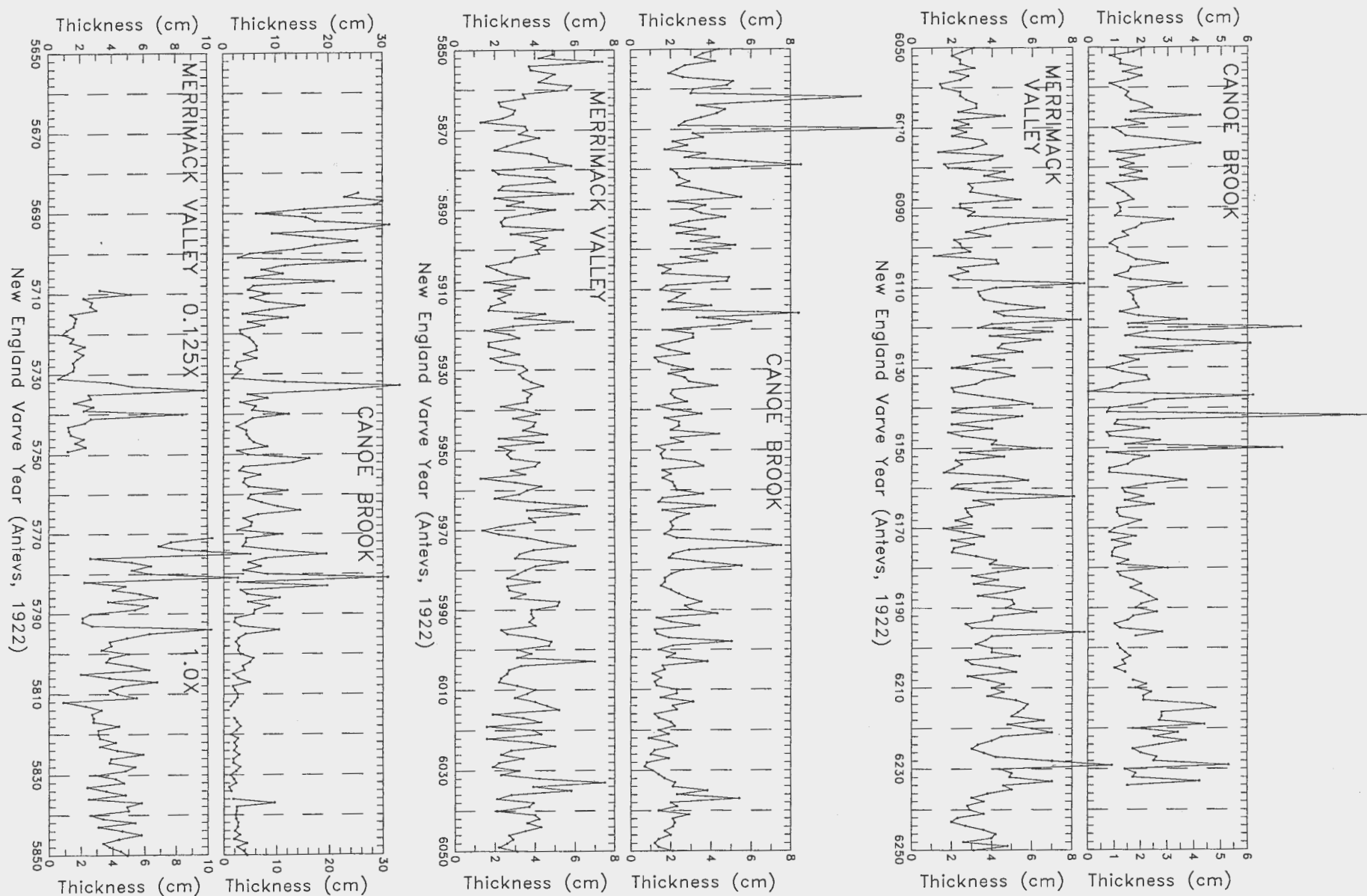


Figure 2-1B. Comparison of varve record at Canoe Brook (top; Ridge and Larsen, 1990) with matching sequence in the Merrimack Valley (bottom; Antevs, 1922). Graphs are 200-yr sections of New England varve years 5650-6250.

section (NE 6150 and 6156; Ridge and Larsen, 1990; Ridge and others, 1999). An additional ^{14}C sample was recovered 300 varve years lower in the section (NE 5858) during a demonstration of coring for Richard Little's video *The Rise and Fall of Lake Hitchcock* in 1998. All of the ^{14}C ages are in agreement with varve counting after correction for the secular variation of atmospheric ^{14}C (see Ridge and others, 1999) except the age in NE varve 6156 which appears to be too old. This age is from redeposited gyttja and peat fragments, which in addition to being pieces of an older eroded organic deposit, likely contain the remains of aquatic plants and algae that are known to produce ^{14}C ages that are too old. All other ^{14}C ages at Canoe Bk. are from terrestrial plant remains. Paleomagnetic (remanence) declination in the varves at Canoe Bk. (Ridge and others, 1999) depict a sharp westward maximum (36° West) that matches an extreme western maximum in declination in correlative sediments in central New York and the Merrimack Valley of eastern New Hampshire. No other western declination maximum of similar magnitude occurred in the northeastern United States between 15 and 10 ^{14}C ka.

Table 2-1. ^{14}C ages from varves at Canoe Brook, Dummerston, Vermont.

Laboratory number	Age $\pm 1\sigma$ (^{14}C yr BP)	$\delta^{13}\text{C}$ (‰)	NE varve number	Material dated	Reference
GX-14781	12,915 \pm 175	-27.1	6156	Peat and gyttja fragments	Ridge and Larsen, 1990
GX-14231	12,355 \pm 75	-27.2	6150	Bulk silt and clay with, non-aquatic twigs and leaves	Ridge and Larsen, 1990
GX-14780	12,455 \pm 360	-27.6	6150	Handpicked non-aquatic leaves and twigs, mostly <i>Dryas</i> and <i>Salix</i>	Ridge and Larsen, 1990
CAMS-2667	12,350 \pm 90	-	6150	<i>Salix</i> twig (AMS)	Norton Miller, pers. comm.
GX-25735	12,660 \pm 50	-28.9	5858	Woody twigs and <i>Dryas</i>	NEW

Road log (cont.) Return to bus on Rt. 5. Head north on Rt. 5 to return to Rt. 91.

- 3.4 (1.7) Turn right off of Rt. 5 to Rt. 91 North (Exit 4 on Rt. 91).
- 11.4 (8.0) Follow Rt. 91 North approximately 8 miles to Exit 6 (north of Bellows Falls). At end of exit ramp turn left onto Rt. 5 North to Rt. 103 and the Williams River.
- 11.8 (0.4) Head west on Rt. 103 into Williams River valley. Rt. 103 follows the surface of deltas and stream terraces graded to the Cold River Stage of Lake Hitchcock.
- 13.4 (1.6) Turn off Rt. 103 onto Parker Hill Road. Leave the bus, which will have to wait at the top of the hill, and walk down the road to the Williams River. The exposure at Stop 2-2 is on the south side of the road just before the bridge crossing the Williams River.

STOP 2-2: Parker Hill Road at Williams River, Rockingham, VT.

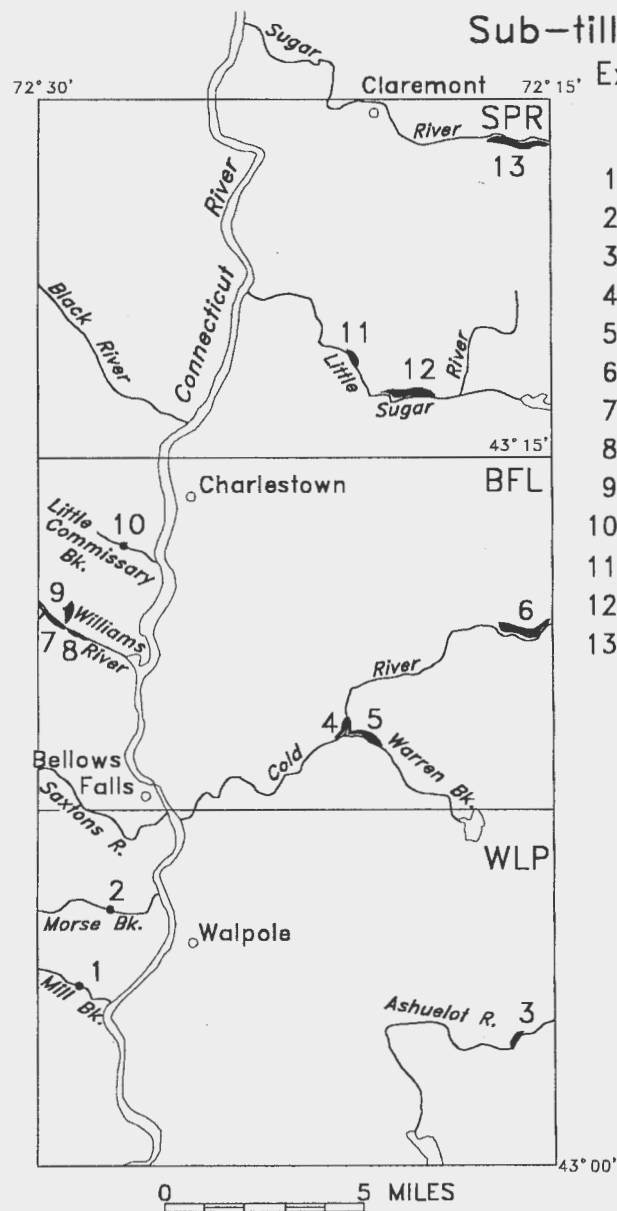
The Parker Hill Road exposure is an example of sub-till lacustrine sediment, which occurs in many east-west trending tributary valleys of the Connecticut River (Figure 2-2). Extensive laminated and sometimes varved lacustrine sand, silt, and clay at these exposures (up to 30 m) may be interbedded with debris flow diamicton and often contain drop sediment. In all cases these deposits appear to represent the impoundment of tributary valleys by the advancing late Wisconsin ice sheet which then eroded and deformed the sequences and buried them beneath thick till. The deposits are not the result of a large basin-wide lake in the Connecticut Valley. They are usually at too high an elevation, fluvial gravel at West Lebanon, NH occurs beneath till at a lower elevation (Larsen, 1987a), and the local northward flow of tributaries favors impoundment during ice advance. So far the base of a sub-till lacustrine sequence has only been seen in the Cold River valley near Alstead, NH (sites 4 and 5, Fig. 2-2) where lacustrine sand and silt was found overlying colluvium and weathered schist (Ridge, 1988). No till has yet been found beneath the lacustrine deposits. This may be a function of preservation locations deep in valleys that were eroded down to bedrock by stream and mass movement activity prior to the last glaciation.

Road log (cont.) Return to bus on Rt. 103 and head back to Rt. 5.

- 15.0 (1.6) Enter Rt. 91 north toward Springfield, VT. This may require us to find a turn-around spot for the bus.
- 22.0 (7.0) At Exit 7 (Rts. 5, 11, 106) get off of Rt. 91 and head toward Charlestown, NH on Rts. 11 East and 5 North.
- 22.5 (0.5) Cross over the toll bridge (\$0.35 for cars) to NH. Take sharp left after crossing bridge following Rt. 11 East toward Rt. 12 North.
- 23.3 (0.8) Road climbs through a massive gravel deposit that is an esker and subaqueous fan complex capped by more recent stream terraces. There are no varves in the valley at this position. After climbing hill to Rt. 12 turn left onto Rt. 12 North.
- 24.2 (0.9) Rt. 12 travels along dune covered stream terrace and rises to the surface of a delta deposited in the Cold River stage of Lake Hitchcock. After climbing a hill take a sharp right turn south onto Old Claremont Road.
- 25.5 (1.3) The Old Claremont Road heads south along the distal topset surface of a Cold River stage delta. Take a sharp left onto Hemlock Road.
- 25.7 (0.2) Pull off to side of Hemlock Road at STOP 2-3.

STOP 2-3: Clay Brook deltas, Charlestown, NH. (Home of Carlton Fisk)

From this point it is possible to see the morphology of two distinct lake levels. Hemlock Road parallels Clay Brook to the south and traverses sandy lake floor deposits that today form the divide between Clay Brook and Beaver Brook to the north. Looking east about 0.3 miles is a



Sub-till Lacustrine Sediment

Exposure: max. obs. ft & sed. type

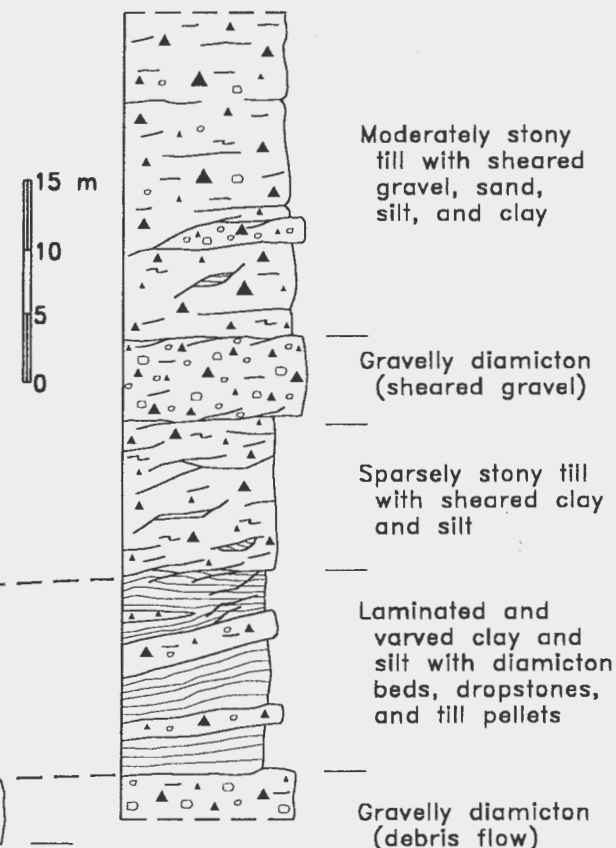
v = varved, m = silt/clay, s = sand,
d = diamicton beds

- 1 Mill Brook: 4md
- 2 Morse Brook: 5m
- 3 Ashuelot River: 6vm
- 4 Alstead, Cold River: 12s
- 5 Warren Brook: 25vms
- 6 S. Acworth, Cold River: 100s
- 7 Williams River: 8sm
- 8 Williams River (STOP 2-2): 30vmd
- 9 Williams River: md
- 10 Little Commissary Bk: 3md
- 11 Little Sugar R: 4vm
- 12 Little Sugar R: 5vm
- 13 Sugar River: 55vmd

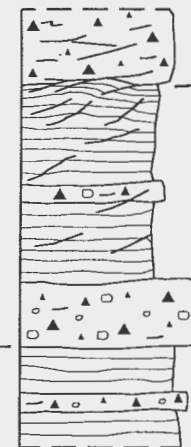
STOP 2-2: Site 8

Williams River, Parker Hill Rd.

Western Gully

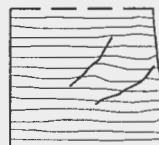


Roadside Exposure



(Road surface)

Eastern Gully



Laminated clay and silt with diamicton beds, dropstones, and till pellets

Figure 2-2. Sub-till lacustrine sediment in the Walpole, Bellows Falls, and Springfield 7.5 x 15' quadrangles and the detailed stratigraphy of the Parker Hill Road section in the Williams River valley (STOP 2-2).

prominent bench composed of bouldery gravel with a farmhouse on its flank (Qcb6 on Figure 2-3). The kettled and partly collapsed surface of this eastern bench reaches an elevation of almost 162 m (531 ft). Qcb6 is an ice-contact delta that can be traced westward along the south side of Clay Brook and into the Connecticut Valley. These features together appear to represent ice-contact deposition in the stable phase of Lake Hitchcock.

Looking west and northwest is also a prominent flat bench, this time composed of pebbly sand to fine sand (Qcd on Figure 2-3). This bench (Old Claremont Rd.) just barely reaches an elevation of 150 m (492 ft) and represents deposition in the Cold River stage of Lake Hitchcock, 10-12 m below the lake's higher stable phase. The Cold River stage is named for a massive delta at the mouth of the Cold River to the south in Walpole, NH. Deltas deposited in the Cold River stage of Lake Hitchcock are the most prominent deltaic features in the Connecticut Valley from Brattleboro northward to Claremont. The town of Claremont sits on the surface of a Cold River stage delta. The Cold River deltas become smaller in size further north probably reflecting a decreasing interval of sedimentation. The Cold River deltas delineate a water plane that is parallel to the stable phase water plane of Koteff and Larsen (1989) and dips at about 0.9 m/km (4.74 ft/mile). The Cold River deltas in southern New Hampshire and Vermont appear to be meteoric because they show no evidence of ice-contact deposition that does occur on deposits at higher elevations. The only known ice-contact deltas of the Cold River stage occur in association with the Littleton-Bethlehem Readvance and later recessional positions in northern New Hampshire (Ridge and others, 1996). Why are the Cold River stage deltas in southern New Hampshire and Vermont so massive as compared to deltas from the stable stage of Lake Hitchcock? It appears that the stable stage did not last as long as the Cold River stage in this section of the valley and most of the stable phase features were dissected by later drainage to the Cold River stage.

The Cold River stage delta at Clay Brook represents all that remains in New Hampshire of the maximum extent of a delta that prograded eastward out of the Black River valley in Vermont. The gentle slope leading up to the flat terrace-like bench to the west is a preserved foreset slope at the front of the delta. The corridor to the north, which extends into the basin in which you are standing, represents the only place where water could pass down the Connecticut Valley when the delta reached its maximum extent. The basin floor is a lake bottom surface in front of the prograding delta. A well in the valley near Clay Brook, about 0.5 miles to the southeast, penetrated 5 ft of gravelly alluvium and 103 ft of fine sand, silt, and clay before bottoming in 6 ft of till. The restricted down valley passage of water at the expanding delta front probably prevented the continued accumulation of deltaic sands and the closure of the valley by delta progradation. A steep scarp cut in till on the east side of the Beaver Brook valley may be the result of erosion by water trying to flow around the delta front.

Road log (cont.) Reverse direction on Hemlock Road and return to Old Claremont Road.

- 25.9 (0.2) Turn left (south) onto Old Claremont Road. The road follows the outer edge of the Cold River stage delta.
- 27.4 (1.5) Follow Old Claremont Road through Snumshire and into Charlestown. Turn left (south) onto Rt. 12 at stop sign in Charlestown.
- 28.0 (0.6) South end of Charlestown. River terraces here are cut into varved clay and silt. Well records in the valley show silt and clay varves overlying sand to a depth of 200 ft. The sand in the base of the valley is the southward extension of the

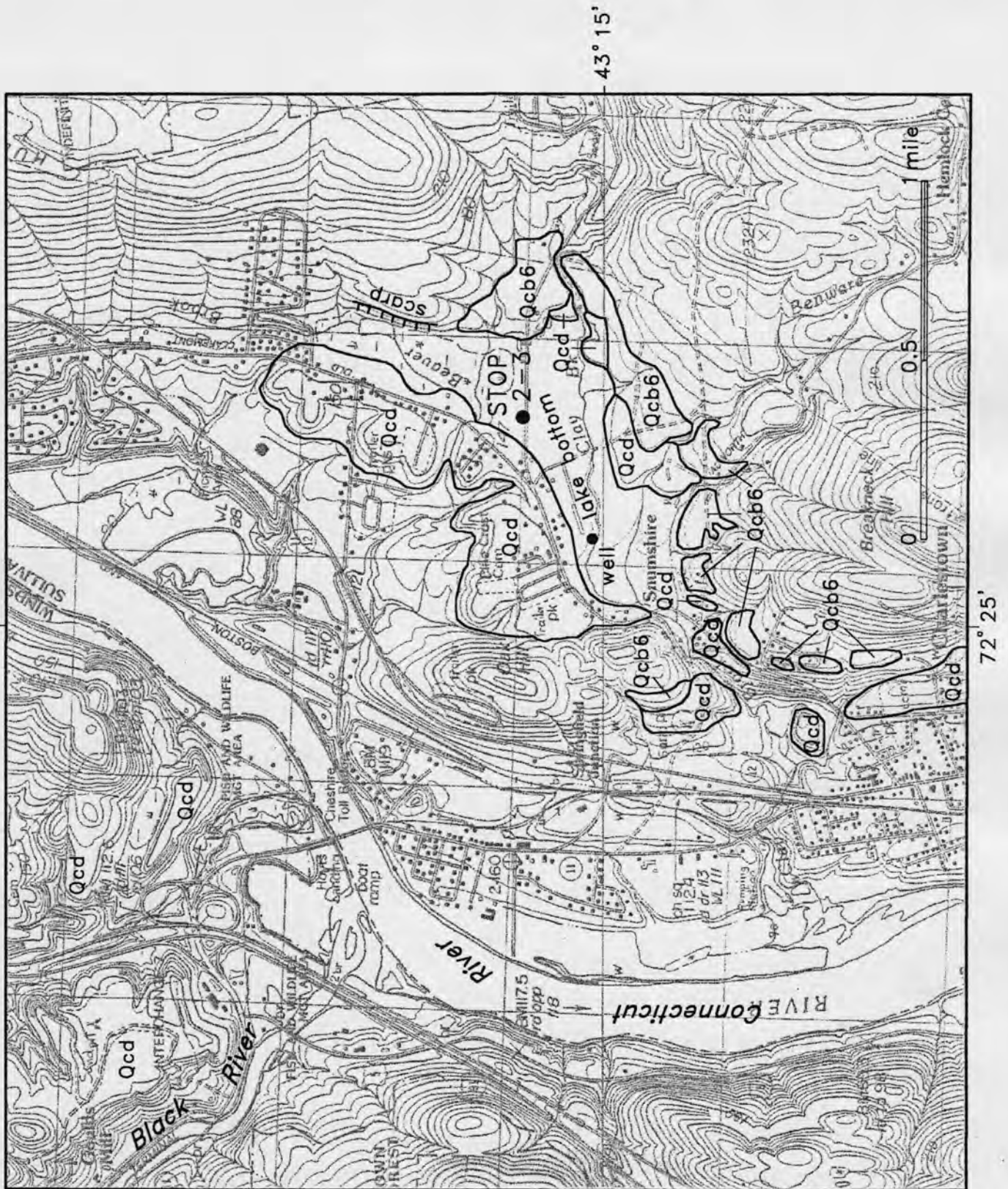


Figure 2-3. Area of Charlestown, NH and STOP 2-3 showing stable phase (Qcb6) and Cold River Stage (Qcd) deltas of Lake Hitchcock. Well record shows 5 ft of gravel over 103 ft of fine sand, silt, and clay over 6 ft of till.

massive esker and subaqueous fan complex that underlies the northern part of Charlestown near the toll bridge. Varves are exposed in the faces of some river terraces. Continue south.

- 30.7 (2.7) Rt. 12A cutoff and overpass. Stay on Rt. 12 South. Road follows bank of Connecticut River, which is impounded by a dam at Bellows Falls.
- 33.4 (2.7) Enter town of North Walpole. The town sits on a river terrace cut into varves that are exposed along the riverbank beneath terrace gravel. Bedrock outcrops also poke upward through the terrace surface in the middle of town.
- 34.3 (0.9) At end of large concrete bridge coming east from Vermont and traffic light turn left (east) staying on Rt. 12 in New Hampshire. Most of the river's water is today funneled through the power plant at Bellows Falls.
- 35.0 (0.7) Rt. 12 follows the natural gorge of the Connecticut River. Turn right (west) onto Vilas Bridge (stone bridge) to cross Connecticut River into Vermont.
- 35.1 (0.05) Immediately after entering Vermont park in lot at fuel oil company. Stop 2-4 is across road from fuel oil company along old railroad grade.

STOP 2-4: Vilas Bridge petroglyphs (LUNCH STOP).

Walk about 20 m down old railroad grade and look over stone wall down onto rock surface in river gorge. The petroglyphs are highlighted with yellow paint. It is possible to climb down the rock face a little further along the railroad grade but this can be treacherous if it is raining. A hundred yards down the railroad grade it is possible to more safely climb down onto the rocks above the river. The Vilas Bridge petroglyphs were originally carved by Native Americans sometime prior to their discovery around 1800. The petroglyphs are cut into ledges of granitic gneiss that is sheared in the base of the gorge. In the 1930's a local chapter of the DAR organized an effort to deepen many of the carvings so they would be more visible. There has been much local speculation about the significance of the petroglyphs including that they commemorate a ceremony or skirmish with settlers in 1775. It has even been proposed that they are Celtic in origin but none of these speculations is supported by facts. The petroglyphs are simple round faces of varying sizes and some of which have one or two horns. According to Dena Dincauze the petroglyphs at Bellows Falls are typical of sites in the northeast, occurring on rock faces near water. The horned faces are typical shaman images that occur at other localities as well. For photos of the petroglyphs and other local historical information refer to www.viz.net/bellowsfalls/.

Road log (cont.) Return trip to Northampton, MA. Continue into Bellows Falls, Vermont crossing canal where water enters the power plant.

- 35.3 (0.2) At intersection in Bellows Falls turn left (south). Follow signs to Rt. 5 South and Rt. 91 South.
- 38.2 (2.9) Turn right onto road leading up to Rt. 91. Follow signs to Rt. 91 South.
- 39.0 (0.8) Enter Rt. 91 South. This is interchange no. 5. Head south to Northampton.
- End of road log for morning trip.**

Road Log continues : Return to The Inn at Northampton, Exit 18, I-91 to pick up vehicles IF necessary. Take Exit 18 on I-91 north to Exit 19, at the bottom of ramp, turn right onto Rte 9. Go 1.8 miles to the intersection with Rte 47. Turn North on Rte 47 (River Drive) and continue north 4.2 miles to Stockbridge Street. Turn right onto Stockbridge Street and then bear left onto Knightly Street where Stockbridge wants to take you right around a curve. Continue on Knightly Street another 0.4 miles. You will need to walk a few hundred yards to the exposure on the farmer's private land. (cumulative mile = 9.4 miles)

Stop 2-5 -- HADLEY SAND DUNE ON LAKE FLOOR ;Excavation into sand dune, north of Knightly Street, Hadley MA; ~701000m E, 4697000m N UTM;Mt. Toby Quadrangle; Tammy Rittenour



Figure 2-5A. Photograph of Hadley transverse sand dune. Cut face is roughly east-west

Exposed in this cut are sediments from a sand dune that formed on exposed lake bottom sediments in the Hadley basin (Figure 2-5A.). This exposure is cut into one of many roughly north-south trending transverse dunes (Figure 2-5B). The sedimentary structures in this dune are very different than those seen in Stop 1-2 (South Hadley Dune). Steeply-dipping foresets are not present, instead the dune is composed of shallowly dipping beds that mimic the gentle lee slope of the dune. A west to south-westerly wind direction is indicated by the sediments and overall morphology of the dune. Sediments in this dune are poorly-sorted, medium to fine-grained sand with some coarse-grained layers. The sediments are poorly sorted. The poor sorting and lack of

well defined foreset beds may indicate that the sediment source for the dune was close by and the dune was not migrating with a grainfall lee slope.

An OSL age obtained from this dune confines the minimum age of complete lake drainage in the Hadley lake basin. The OSL age suggests that the Hadley basin completely drained prior to the deposition of the sand dune at 14.3 ± 1.6 cal kyr BP (12.3

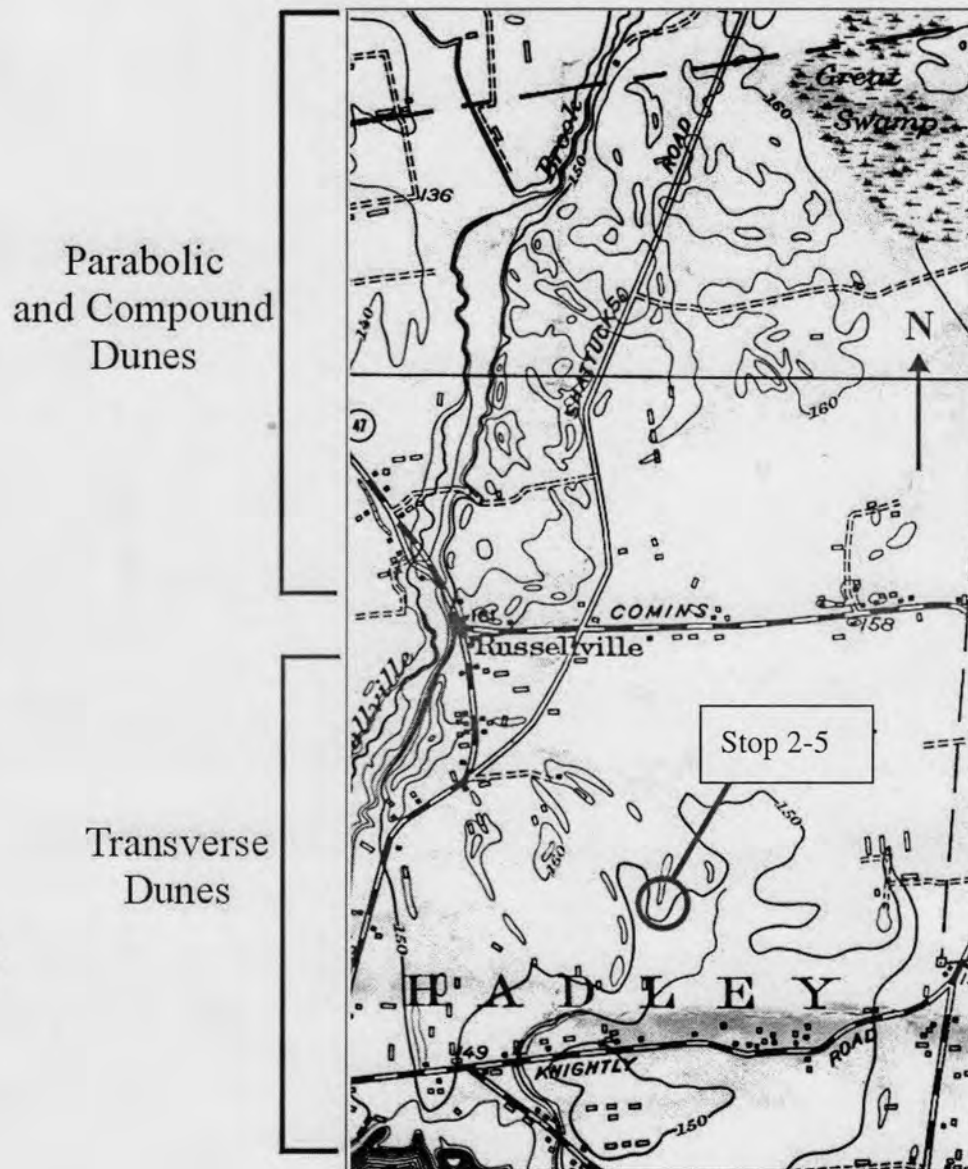


Figure 2-5B. Topographic map with the location and orientation of the Hadley transverse sand dune indicated. The sand dunes in the southern half of the figure are mostly transverse dunes while those in the northern half are parabolic and compound dune forms. Contour interval is 10 feet, Mount Toby Quadrangle.

¹⁴C kyr BP). Radiocarbon ages from the UMass core indicate that varve deposition continued within the Hadley basin until 14.0 cal. kyr BP (12.0 ¹⁴C kyr BP). These age estimates constrain the complete drainage of Glacial Lake Hitchcock in Massachusetts to between 14.0 and 14.3 cal kyr BP.

Road Log CONTINUES: Return to the Inn at Northampton by reversing the route. Turn left out of the Inn at Northampton onto Rte 5&10 and drive north into Northampton on King Street. In downtown Northampton, take a left onto Rte 9 (Bridge Street) and continue through two a few blocks to left turn onto Rte 66 in front of Smith College. Take Rte 66 (variously labeled Chapel Road and Rocky Hill Road) 2.5 miles to intersection with Florence Road. Take a left and then an immediate right onto the continuation of Rocky Hill Road. Go about 1 mile and park along side of road. Hike down gas-pipeline rightaway about 200 yds. (cumulative miles= 4.9 miles)

STOP 2-6 EASTHAMPTON/NORTHAMPTON PINGO SCARS; Pingo Scars filled with water in woods; South of Park Hill Road, West of Bassett Brook, Northampton MA; ~690000m E, 4684500m N UTM; Easthampton Quadrangle; Tammy Rittenour

At this site there are 4 closed depressions that we are interpreting as pingo scars. They are semi-circular depressions with very slight up-raised rims that are clustered together on the lake bottom surface near the 200 ft contour interval (61 m). Several other clusters of similar closed depressions are found throughout this region and they all lie at the same elevation (210 – 190 ft asl) (Figure 2-6A). The morphology of the closed depressions in addition to their occurrence in clusters at the same elevation provides strong evidence that these were open-system pingos that formed after lake drainage. Interestingly the elevation that these pingo scars formed is the approximate elevation of (or just below) the water level of the lower-level Lake Hadley. These pingo scars developed in the silts and sands that were washed into Lake Hadley and deposited over the underlying varves at the site. This stratigraphy of water saturated silts and sands overlying varves may have promoted the development of open-system pingos along the edge of Lake Hadley after a minor drop in lake level.

Road Log: continues: Reverse Route to the Inn at Northampton.

End of Trip. Have a safe drive home!

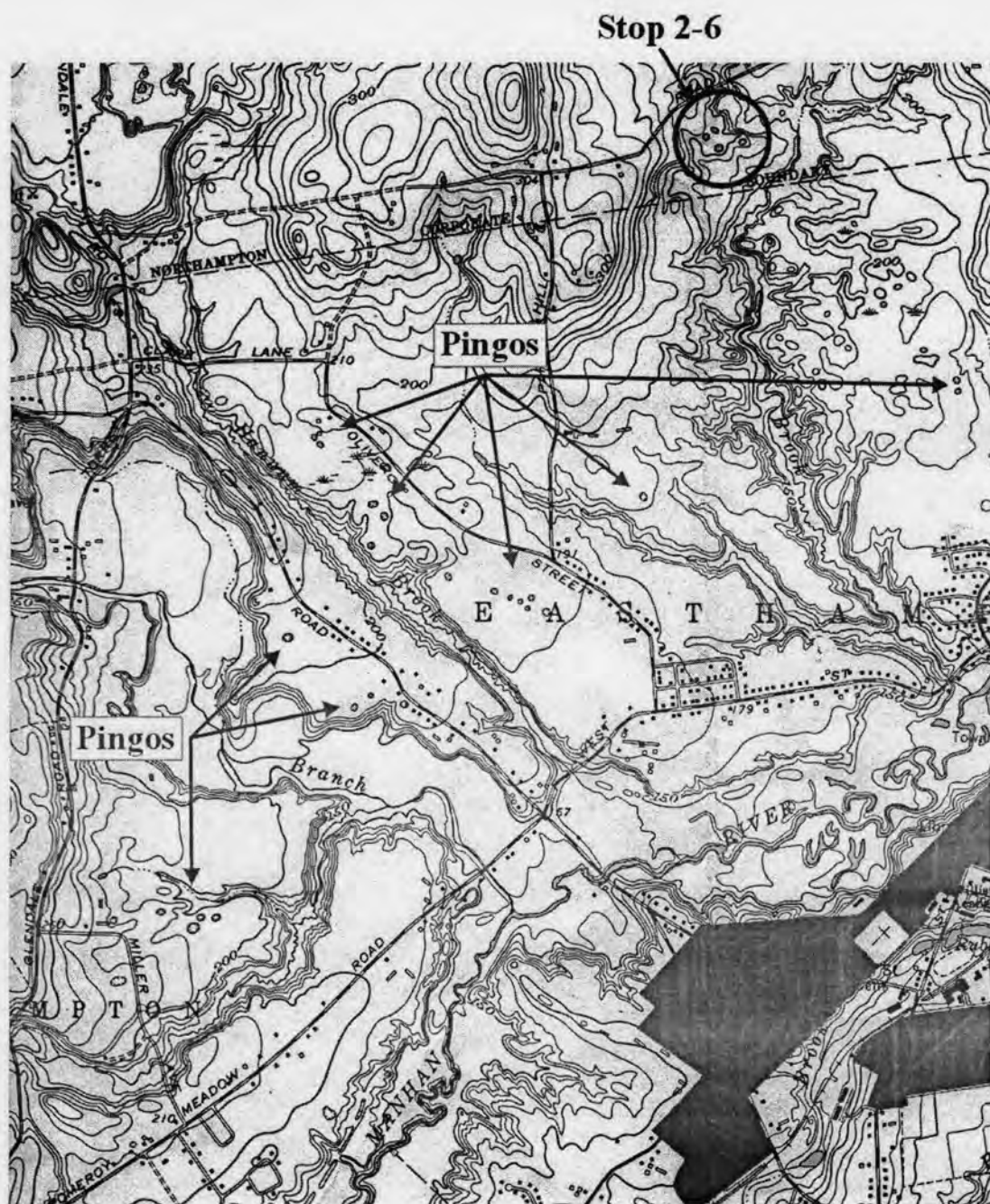


Figure 2-6A. Topographic map of the area between Northampton and Easthampton with pingo scar locations indicated. Contour interval is 10 feet, Easthampton Quadrangle.

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