ICE-WEDGE CASTS, PINGO SCARS, AND THE DRAINAGE OF GLACIAL LAKE HITCHCOCK

by

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(to whom we dedicate this fieldtrip)

INTRODUCTION

A harsh postglacial climate, in which permafrost was present, existed for several thousand years following deglaciation until after the drainage of glacial Lake Hitchcock. On this fieldtrip (fig. 1) we will examine paleoclimatic evidence in the form of ice-wedge casts, pingo scars, eolian deposits, cryoturbation structures, and the paleobotanical record, that provides evidence for this harsh climate as well as new evidence concerning the timing of the drainage of Lake Hitchcock.

Wedge-shaped structures developed in glacial meltwater deposits capped by eolian mantle have been exposed for many years at the Hain Brothers pit in Windham CT. We will examine evidence (including their demonstrable polygonal ground pattern) which argues strongly for the interpretation of these structures as ice-wedge casts indicative of permafrost conditions. We will examine the stratigraphy of the glacial meltwater deposits at depth in the section, the deformational characteristics of the linear wedge structures that cut the surface of these deposits, and the "pedogenic soil" (Agawam series) developed on the cryoturbated eolian material that mantles the surface and penetrates the wedge structures. Evidence will be presented that argues against former interpretations of the structures as 1) "pseudo-ice-wedge casts" resulting from tensional faulting unrelated to permafrost (Black, 1983) and 2) "earthquake-induced fissures" that served as conduits for liquified sediment (Thorson and others, 1986).

On lake-bottom surfaces of glacial Lake Hitchcock we will visit several localities where clusters of rimmed depressions are interpreted as pingo scars that formed as a result of the growth and decay of permafrost lens ice on the lake-bottom surface following drainage of the lake (Stone and Ashley, 1989; Stone and others, 1991). These features occur extensively on surfaces below paleo-lake level that were not fluvially terraced by the postlake stream incision. We will visit the Kelsey-Ferguson clay pit, excavated in a terraced lake-bottom surface that has no pingo scars but does exhibit well-developed eolian morphology in the form of dunes and deflation hollows.

We will present and discuss new ^{14}C dates and plant macrofossil and pollen assemblages from the upper section of Lake Hitchcock lake-bottom sediments and from the clastic and organic fill of the pingo-scar depressions. Dates from the pingo-scar peat clearly demonstrate drainage of glacial Lake Hitchcock at least 2000 years before the 10,700-yr BP date of Flint (1956), and dates on detrital plant debris in the upper lake beds indicate that the lake probably drained nearly 2000 years before the suggested 12,400-yr BP date of Ridge and Larsen (1990).

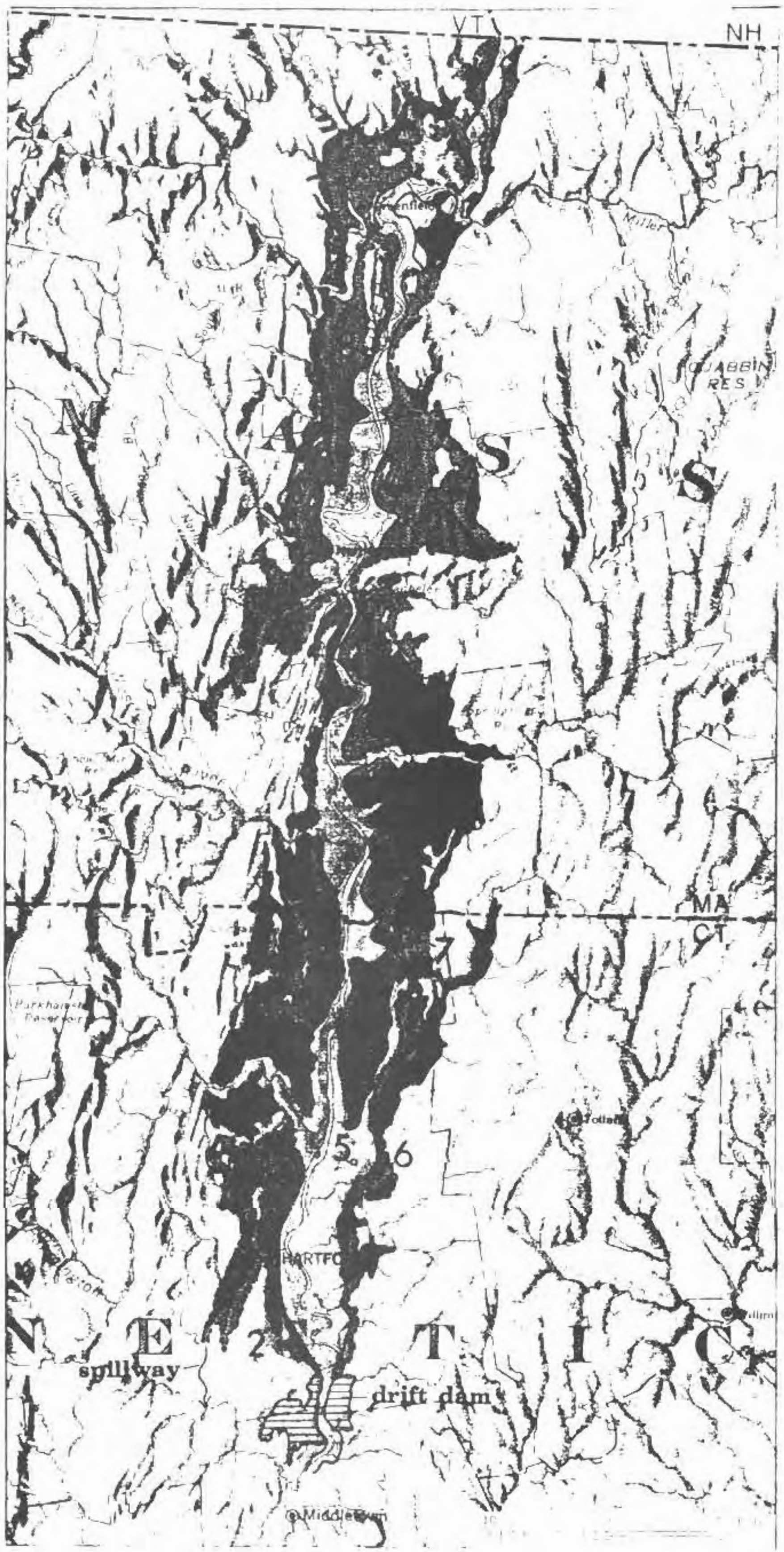

Figure 1. Glacial Lake Hitchcock surfaces in Connecticut and Massachusetts. Pingo scars occur extensively on the lake-bottom surfaces. Numbered localities are FIELDTRIP STOPS.

delta surfaces — lake-bottom surfaces — post-lake incised terrace surfaces

ICE-WEDGE CASTS AND POLYGONAL PATTERNED GROUND

Wedge-shaped structures that deform late-Wisconsinan glacial sediments have been found, mostly in solitary occurrence and without demonstrable polygonal ground pattern, throughout southern New England and have been interpreted by a number of workers (fig. 2) as ice-wedge casts indicative of former permafrost conditions. The wedge structures were first suggested to be ice-wedge casts by J.P. Schafer (Schafer and Hartshorn, 1965; Schafer, 1968). For many years the best exposure of multiple structures has been in the Hain Pit in Windham CT, where a number of workers have both described and interpreted the structures in different ways. They are clearly linear, vertical, filled extensional fractures that deform the upper few meters of an eolian-mantled glacial meltwater deposit.

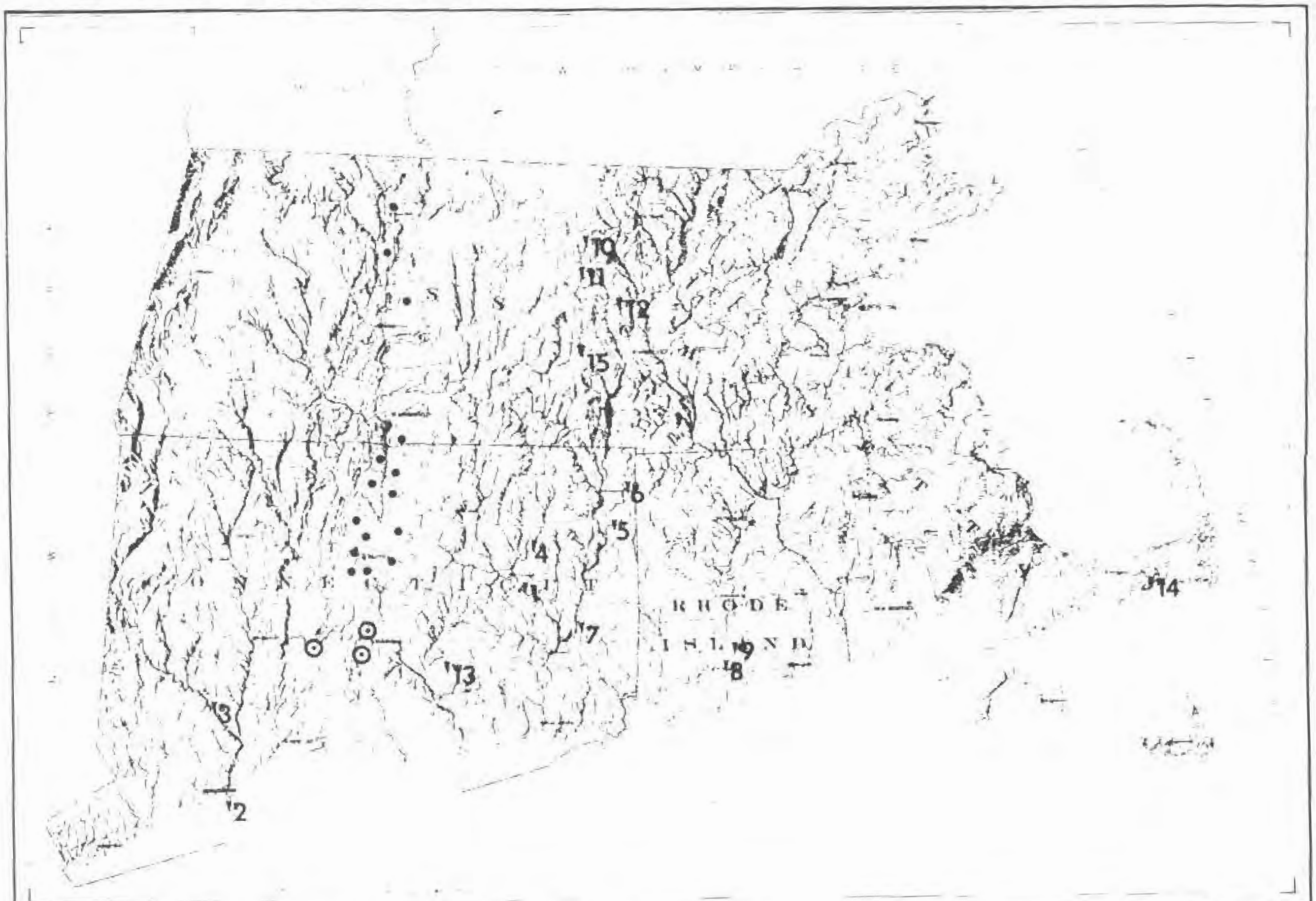

Figure 2. Locations of features indicative of former permafrost in southern New England.

Ice wedge casts- Number keyed to references below

1. Hain Pit, South Windham, CT (this paper)
2. Lordship, Stratford, CT (Denny, 1936)
3. Seymour, CT (Schafer, 1968)
4. Bedlam Corner, Chaplin, CT (Schafer, unpublished)
5. Dayville, CT (E.H. London, unpublished)
6. Munyon Road, Thompson, CT (Schafer and Hartshorn, 1965)
7. Jewett City, CT (B.D.Stone, 1978)
8. University Pit, Kingston, RI (Hartshorn and Schafer, 1965)
9. W. Allentown Rd., N. Kingston, RI (Boothroyd and Lawson, 1989)
10. Hubbardston, MA (J.R. Stone, unpublished)
11. Barre, MA (Larsen, 1979)
12. Holden, MA (B.D.Stone, unpublished)
13. Moodus, CT (O'Leary, 1975)
14. Dennis, MA (Oldale, 1974)
15. E. Brookfield, MA (Stone and others, this volume)

- **Pingo-scar clusters**- on drained Lake Hitchcock surfaces

⊙ **Pingo-scar clusters**- on other glacial lake surfaces

extensively

rfaces

In cross-section (fig. 3), the structures are vertical and wedge-shaped, up to 1.0 m wide at the top and tapering downward to termination commonly at depths of 3-4 m. Deformed wall strata attenuate downward in steeply nested V's and become (by gradual loss of stratification) the nonbedded central cores of the filling. The cores are characterized by vertically oriented clasts. The upper part of filling consists of tongues or plugs of eolian sandy silt containing ventifacted pebbles. At its base this material is often gray (nonoxidized) and displays crude concave-upward layering. The eolian material commonly penetrates the upper 1-2 m of the wedge; it penetrates as deep as 3 m in places along an east-west striking feature currently being excavated at the north end of the Hain pit. Where the deformation cuts across wall beds of medium to coarse sand, rather than gravel, normal faults bound the fracture and the sand may be deformed downward into the filling by a series of normal faults. In some cross-sectional exposures, particularly in the sandy wall-bed material, thickened and/or upturned bedding and reverse faults are present on both sides of the normally downdropped fill. This indicates initial pressure outward from the center before inward collapse. The above cross-section description and (fig. 3) sketch apply to views where the pit face is vertical and perpendicular to the strike of the wedge. It is important to note that in situations where this is not the case, the wedge structure will appear to be asymetric and non-vertical. In situations where the pit face is oblique or parallel with the wedge strike, and in places where the pit face cuts wedge intersections, complex configurations occur which are easy to misinterpret.

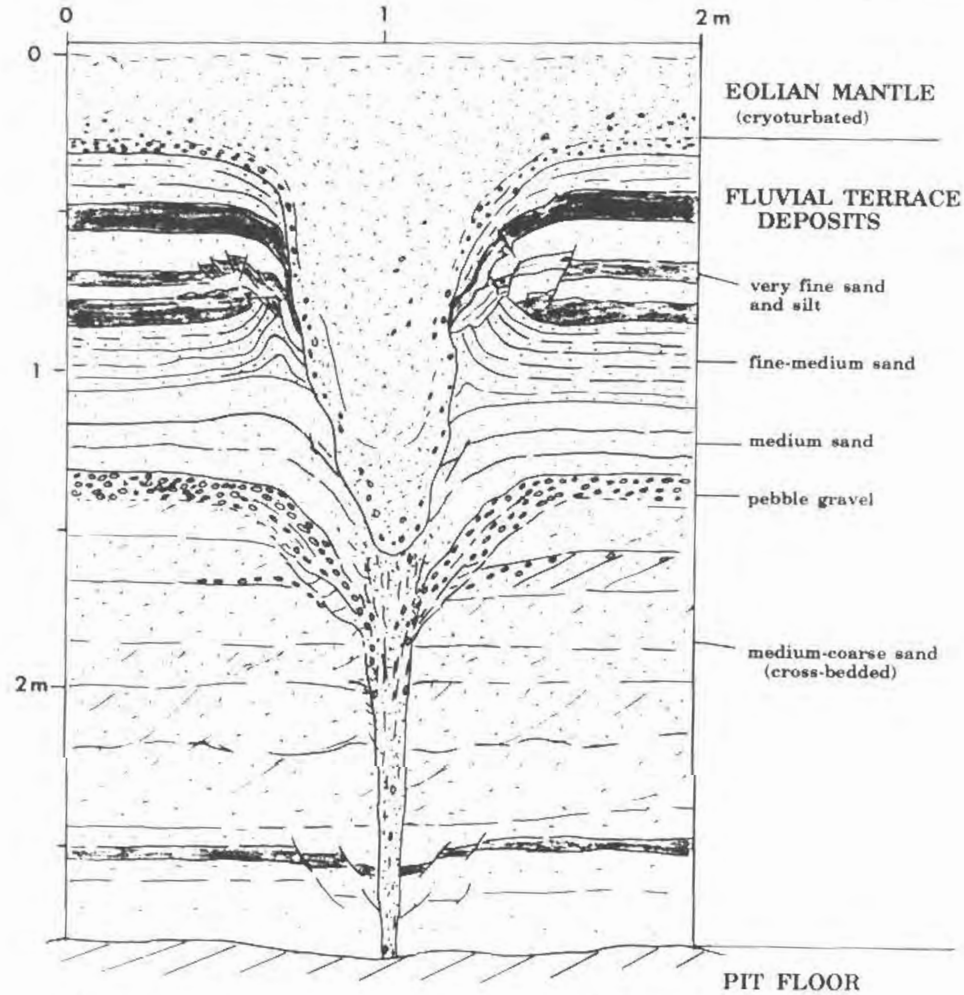

Figure 3. Sketch of ice-wedge cast, Hain Pit, Windham, CT.

In plan view (fig. 4), the wedge structures are linear, filled fractures that locally intersect. Figure 4E (from Oakes, 1992) shows all recorded strike directions for wedges in the Hain Pit including those measured by Thorson and others (1986). The nature of their horizontal pattern has been seen in a limited way by recording wedge traces through continuing excavation of pit faces (Thorson and others, 1986) and at depth by traces on small areas of the current pit floor.

On some aerial photographs of the pit area, traces of the wedges can be seen on surfaces stripped of topsoil prior to excavation. Only small surface areas in this condition have been observed, thus resulting in only an incomplete and speculative assessment of the size and character of the polygons bounded by the wedge fractures. Recently, close examination of 1980 airphotos (FL-69, 4198-99) in the Hain pit area has revealed a polygonal ground pattern on a 250-m by 180-m field located 200 m north of the Hain Pit (fig. 4A and B). At the time these photos were taken (April 19, 1980) this particular field had optimum conditións to allow higher

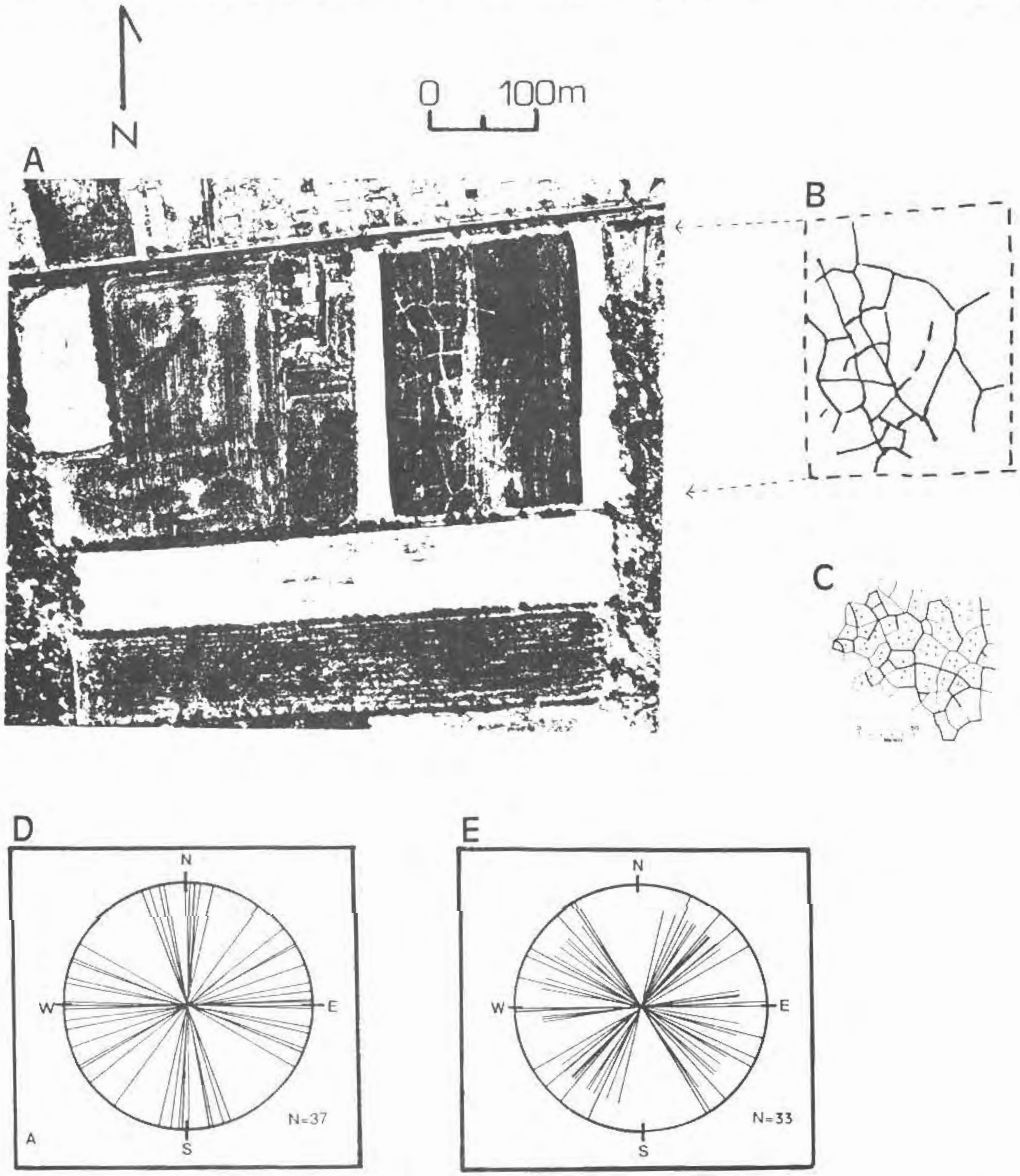

Figure 4. A) Section of airphoto (April 1980, FL-69, 4198-99) showing polygonal ground pattern on field north of Hain Pit, Windham, CT. **B)** Ground pattern traced directly from original airphoto. **C)** Polygonal ground pattern on sand and gravel surface in Barrow, Alaska (from Black, 1974, fig. 23). **D)** Strike directions measured from patterned ground on field north of Hain Pit (from Oakes, 1992). **E)** Wedge-strike directions recorded from Hain Pit (from Oakes, 1992).

moisture content in the finer-grained eolian material of the upper wedge fill to be visible from the air. Similar conditions allow relict patterned ground in central Illinois to be visible on airphotos (Johnson, 1990). Patterned-ground polygons on the field north of the Hain pit have diameters of 20-50 m; these polygons appear to be secondary to a larger, 150-180-m-diameter polygon. Strike directions on the field pattern plotted from airphoto measurements are shown in figure 4D (from Oakes, 1992). One can compare the configuration and size of polygons with those on a surface underlain by sand and gravel at Barrow, Alaska (fig. 4C) from Black (1974, fig. 23).

The Pseudo-Ice-Wedge Cast Interpretation

The strongest opponent of permafrost genesis of wedge-shaped structures in New England glacial deposits was Robert F. Black. Black argued that the postulated thermal environment during deglaciation of Connecticut (i.e. wet-based nature of the ice sheet, rapidity of its retreat, nearness to the ocean, and latitude) precluded the possibility that permafrost could have existed for 1,000 yrs or more during deglaciation (Black, 1983). Black described the wedge structures at the Hain Pit and the Bedlam Corner pit 12 km to the north as "pseudo-ice-wedge-casts because they are deceptively similar to true ice-wedge casts". He argued that these features did not meet the criteria as defined by Black (1976) to be interpreted as permafrost features. His reasons included; 1) commonly solitary nature of wedge occurrence, 2) lack of polygonal ground pattern with appropriate polygon diameters for the host material, 3) absence of pressure effects and upturned wall beds in at least some of the structures within a group, and 4) absence of other permafrost indicators within the region. Black also described what he believed to be evidence for relatively recent deformation associated with these structures because at least one of them appeared to have B-horizon soil material deep within the wedge filling (he cites Harvey Luce, oral comm., 1979), and because many of the wedge structures cut (and therefore postdate) horizontal iron-stained layers. Black concluded that the wedge structures in Connecticut are most likely tensional fractures produced by the loading of coarse clastics on fine clastics, and are associated with a lowering water table produced by postglacial stream incision; he concluded that deformation continued into modern times.

The Earthquake Interpretation

Interestingly, sediment deformation produced by permafrost may resemble that produced by earthquakes. Thorson and others (1986) interpreted the wedge structures at the Hain pit as earthquake-induced ground fissures. They describe the sediment filling, in particular the upper plug material of massive, gray silty sand, as subjacent lacustrine material that liquified during seismic excitation, was vented to the surface, and then flowed into the upper part of the ground fissures. Thorson and others (1986) also believed that the structures were significantly younger than the glacial sediments in which they occur; they point out, as did Black (1983), that the wedge structures cut strongly iron-stained layers that result from prolonged contact with aerated ground water to produce the oxidation. They suggest that the seismic event which produced the fissures and liquified sediment fill was clearly prehistoric but may have occurred within the past few thousand years. A 1050±100-yr BP date on slightly organic silty colluvium associated with the surface of one structure is used to substantiate the recent age of these features; however, the context and interpretation of this date, as presented, are extremely nebulous.

The Ice-Wedge Cast Interpretation

The wedge-shaped structures at the Hain pit and patterned ground on the field to the north are here interpreted as ice-wedge casts indicative of permafrost conditions in eastern Connecticut following retreat of the ice sheet from that area about 16.5 ka. These features can now be shown to possess all of the requisite characteristics identified by Black (1976) and other authorities on past and present permafrost to be classified as ice-wedge casts. These characteristics include:

1) similarity to ice-wedge casts found in northern latitudes today in areas of degrading permafrost
2) large-diameter polygonal ground pattern (20-50 m), which is consistent with the thermal expansion properties of sand and gravel
3) upturned wall beds and pressure effects preserved within wall-bed sediment in some of the structures which provide evidence for ice expansion

4) gently downwarped wall beds in nested V-shaped layers that are evidence for slow grain-by-grain replacement of ice with sediment in the fracture
5) upper part of wedge filling consists of downdropped eolian material that commonly contains ventifacted pebbles and displays crude concave upward layering that suggests filling of an open contraction crack within the seasonally thawed active zone, and
6) presence of complex soil involutions clearly formed by cryogenic processes under severe cold-climate conditions although not necessarily in permafrost; these commonly "pot-shaped" cryoturbation structures occur separately from the wedges, but also in places deform the wedge structures.

None of the many wedge structures currently exposed in eastern Connecticut involve deformation of the modern soil horizon. Dr. Harvey Luce (professor of soil science, University of Connecticut) has examined many wedge fillings at the Hain pit as well as the normal soil profile (see STOP 1, section f). He describes the pedogenic soil as Agawam series developed in relatively thick (up to 40 cm) eolian mantle overlying well-drained sand and gravel. The normal soil profile crosses the wedge features therefore, soil formation postdates the wedge deformation. Strongly oxidized layers in the host sands and gravels appear to be cut by the wedge structures and associated deformation. Prolonged exposure to aerated ground water is required to produce the oxidation, as stipulated by Black, (1983) and Thorson and others, (1986). However, the oxidation takes place in the coarser-grained areas where permeability is high regardless of the bedding configuration of those areas. This point is demonstrated by the oxidation within the coarsest material in complex sedimentary structures in these same beds. The oxidation postdates the sedimentary structures and the wedge deformation as well.

Another requisite factor defined by Black (1976) for the interpretation of wedge structures as ice-wedge casts is that there must be other cold-climate indicators in the region. Such evidence is present, in the form of pingo scars, extensive eolian deposits, and tundra vegetation, and is discussed below.

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 resemblence 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/km^2. 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

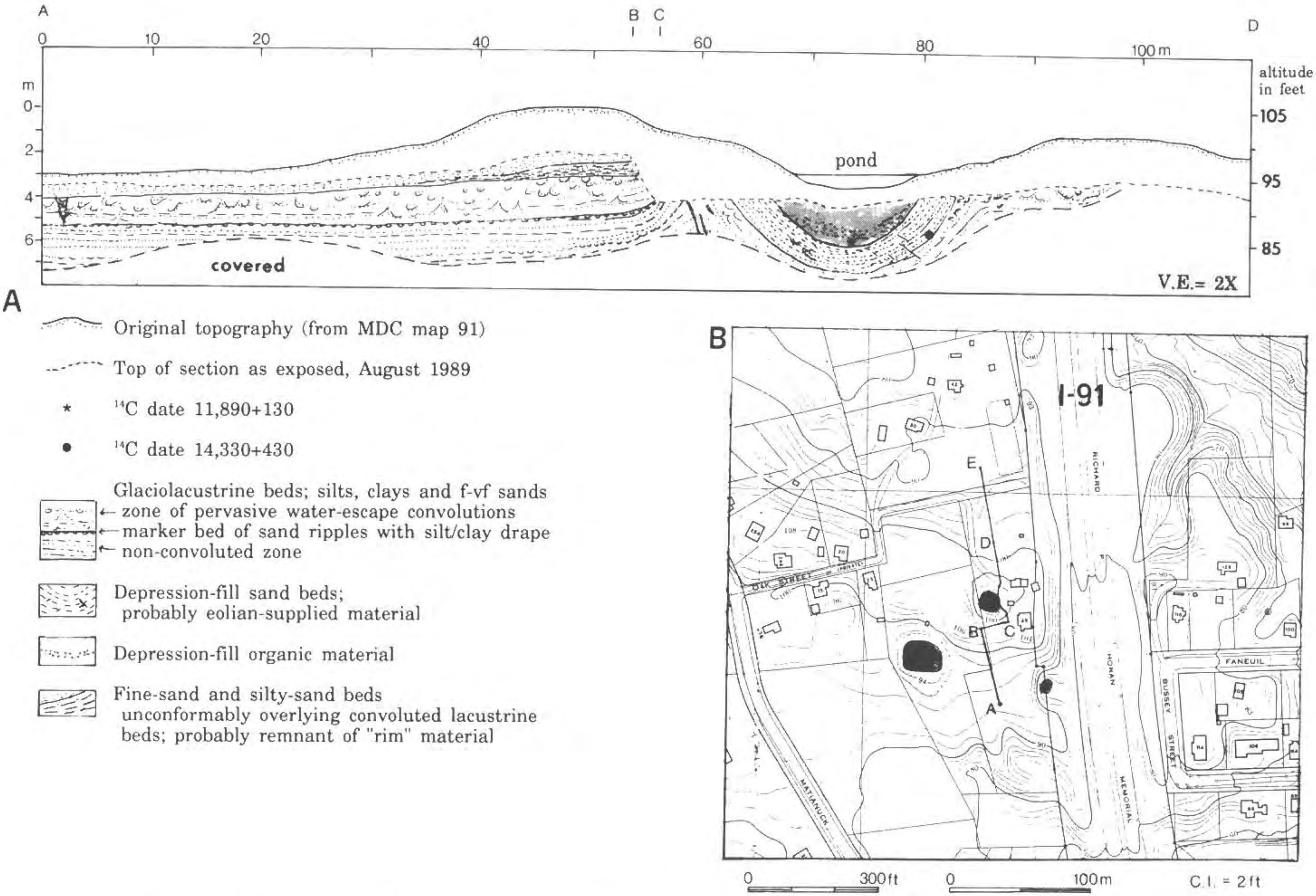

Figure 5. **A)** Measured Section at I-91 roadcut north of Hartford CT. **B)** Section of MDC Map 91 (1977 ed.) showing topography of area and location of section line (A - E).

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.

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±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: <u>Twigs</u>- *Vaccinium oxycoccos* (small cranberry), *Vaccinium uliginosum* (bilberry); <u>Seeds</u>- *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±130 yr BP** (Beta-34820). A second (AMS) date of **12,050±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: <u>Needles</u>- *Abies balsamea* (balsam fir), *Picea sp* (spruce), cf *Juniperus* (juniper), *Larix* (larch); <u>Seeds</u>- *Picea sp* (spruce); <u>Bracts</u>- *Abies balsamea* (balsam fir), *Betula sp* (birch); <u>Leaves</u>- *Vaccinium sp* (bilberry or cranberry); <u>Twigs</u>- *Larix* (larch), *Betula sp* (birch), Ericaceae- heath family, *Myrica gale* (sweet gale), *Shepherdia canadensis* (soapberry); <u>Other small plants</u>- *Gaultheria procumbens* (mountain tea), *Vaccinium oxycoccos* (cranberry), *Cyperaceae* (sedge); <u>Other plant parts</u>- Nymphaceae (rhizome fragments-water lilly), *Equisetum sp* (stems-horsetail), Gramineae- stems grass sp, *Chara fragilis* (green algae), *Mycorrhizal sclerotia* (fungi); <u>Animal</u>- beetle exoskeleton fragments, *Daphnia* (water flea); <u>Pollen</u>- 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±240 yr BP** (Beta-46514); the base of the continuous peat body yielded a date of **12,350±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±100 yr BP** (GX-17050) was obtained from core F-5; a date of **7,850±220 yr BP** (GX-17053) from core F-3.

A date of **12,200±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.

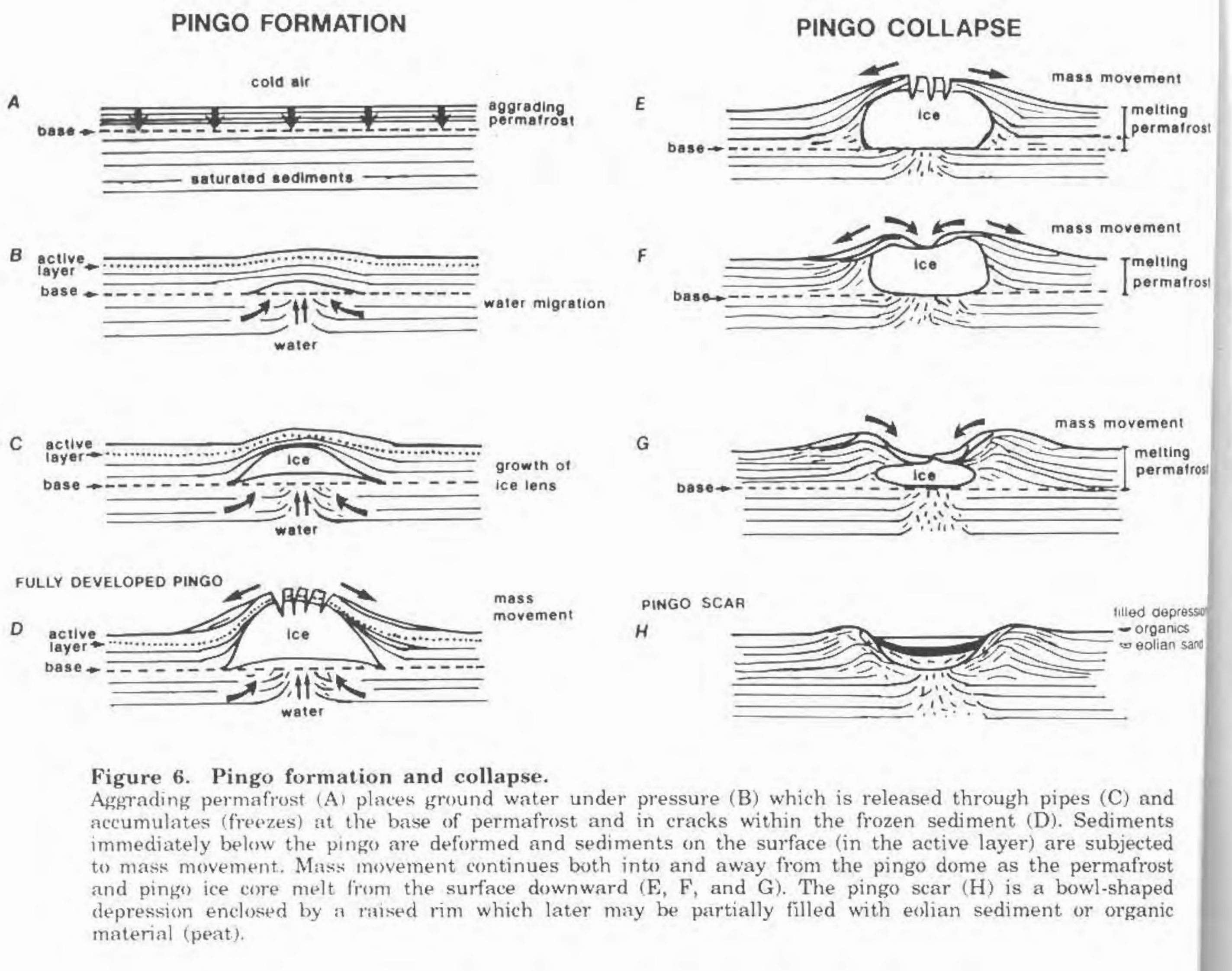

Figure 6. Pingo formation and collapse.
Aggrading permafrost (A) places ground water under pressure (B) which is released through pipes (C) and accumulates (freezes) at the base of permafrost and in cracks within the frozen sediment (D). Sediments immediately below the pingo are deformed and sediments 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 Penninsula, 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 >12.0-12.6 ka from basal organic fill of the rimmed depressions on the Hitchcock lake bottom clearly demonstate 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.

A preliminary account of plant fossils from Matianuck Ave. site
by Norton G. Miller, Biological Survey, New York State Museum, Albany, NY 12230

Lake-bottom sediments from the Matianuck Ave. site in Windsor CT contained abundant pollen and plant macrofossils. Two samples, mostly lacking large plant fossils and each limited to a single bed, one 5 cm below the other, were taken for pollen analysis from a clean face of a 20x20x20-cm block of sediment. The two samples were separated by four thin sandy beds. The samples were processed in the laboratory following standard procedures (HCl and HF digestion of inorganics, acetolysis, floatation in $ZnCL_2$). Both pollen residues contained large amounts of pollen of herbs, over 50% of a sum (percentage base) that included all terrestrial plants (trees, shrubs, and herbs; with pteridophytes, bryophytes, and unknowns and degraded pollen excluded from the sum).

The pollen spectra are unusual in having high percentages of grass (Gramineae) pollen (41%), a relatively modest representation of sedge pollen (14%, 16%), and the occurrence of pollen of many kinds of herbs and shrubs, including *Armeria* (0.4%, 0.6%); *Dryas* (0.3%, 1.4%); *Oxyria* (0.8%, 3%); *Saxifraga* (0.2%, 0.3%); *Thalictrum* (2%, 4%); undifferentiated Caryophyllaceae (4%, 4%), Compositae (2%, 3%; excluding *Artemisia* (0.4%, 0.2%)); Cruciferae (0.1%, 0.3%), Ericaceae (0.2%, 0.5%) and Rosaceae (0.1%, 0.1%); *Salix* (4%, 5%), and *Alnus* (including *A. crispa*). Pollen of aquatics was limited to single grains of *Hippuris* and *Nymphaea*.

Tree pollen percentages were 29% and 33% of the sum, with pine (*Pinus*) the predominant type (14%, 19%). Much of the pine pollen was of the jack pine (*P. banksiana*) type. Spruce (*Picea*) percentages were low (4%, 5%), and small amounts of fir (*Abies*) pollen were present (0.1%, 0.6%). Pollen from hemlock (*Tsuga*), beech (*Fagus*), sugar maple (*Acer saccharum*), ironwood (*Carpinus* and/or *Ostrya*), elm (*Ulmus*), ash (*Fraxinus*), basswood (*Tilia*), hickory (*Carya*), black walnut or butternut (*Juglans*), and aspen or poplar (*Populus*) occurred in trace amounts (generally less than 0.2%). Oak (*Quercus*) and birch (*Betula*) pollen was more frequent, 1-2% and 1% of the sum, respectively).

These pollen assemblages indicate an essentially treeless vegetation. The tree pollen present (boreal conifers, spruce, and jack pine) probably came from distant (i.e. not local) sources. Some of the other pollen in the sediment may have been deposited by wind, but most of it is probably of terrestrial detrital origin, deposited in Lake Hitchcock along with the plant detritus. Therefore, direct comparisons with the pollen spectra of the same age from small lakes are problematic. The terrestrial detrital origin of much of the pollen may explain why grass is so abundant, whereas it is poorly represented in the basal late-glacial silts and clays of small lakes basins in the New England-New York region.

Plant macrofossils were abundant in the sediment; these were disaggregated by swirling a sample in a dish of water until the individual leaves and other fossils were freed from one another. Inorganic sediment in suspension was decanted as required. In a second treatment, bulk sediment samples were washed free of small inorganics in nested 500-um- and 250-um-mesh sieves. The organic-rich residues were transferred to jars prior to examination under a dissecting microscope. To help free the fossils from adhering silt and clay, the samples were put in beakers, flooded with 5% Na_2CO_2 and heated to 50^0 C on a hot plate for 3 hr (with periodic stirring). To help keep the fossils intact, vigorous agitation of the sample was avoided.

The plant macrofossil assemblages consisted of leaves, twigs, bud scales, fruits, and seeds of vascular plants and fragments of leafy moss plants. The assemblages have not been studied completely, but some of the species already identified include the dwarf willow, *Salix herbacea* L. (leaves of which are the most abundant plant macrofossil in the organic debris); alpine meadowrue (*Thalictrum alpinum* L.), *Sibbaldia procumbens* L., and *Oxyria digyna* L. (which are represented by fruits); *Armeria maritima* L. (represented by calyces); and the moss *Aulacomnium turgidum* (Wahlenb.) Schwaegr. Some of these macrofossils are illustrated in figure 7. In addition, fruits of sedges, *Potentilla* and Compositae and leaves of other species of willow and of blueberries (*Vaccinium*) are present, as are about 20 additional species of mosses.

The contemporary ranges of these species of seed plants and the moss are largely arctic-alpine, with southernmost occurrences in northeastern North America associated with the highest mountains in this region (e.g. the Adirondack High Peaks, (New York), the Presidential Range (White Mountains, NH), the Shickshock Mountains (Gaspe Peninsula, Quebec), and alpine areas of Newfoundland and Labrador). *Salix herbacea* and *Aulacomnium turgidum*, for example, now grow in the alpine tundra of the Adirondacks, White Mountains, Mt. Katahdin, and in appropriate places northward, but *Thalictrum alpinum* occurs no farther south at present than in suitable habitats on the Gaspe Peninsula. Thus, these species do not presently share identical ranges but are similar in being arctic and alpine in distribution and occurring in areas that are treeless.

The pollen and plant macrofossils indicate a preforest stage in the late-glacial vegetation of New England. They are evidence of herb- and shrub-dominated plant communities on surfaces above the shore of Lake Hitchcock. The communities comprised species that today are common in subarctic and arctic Canada; they occur only sporadically in the northeastern United States and adjacent parts of Canada in edaphically specialized sites on cliffs that are microclimatically suited to arctic and alpine plants. The plant fossils at this site extend the record for such vegetation in the Connecticut River Valley 325 km south of the Columbia Bridge site, near Colebrook, NH (Miller and Thompson, 1979). At Columbia Bridge, a boreal plant assemblage is dated at 11.4-11.5 ka. This assemblage is present 2,500 years earlier at the Windsor, CT, site.

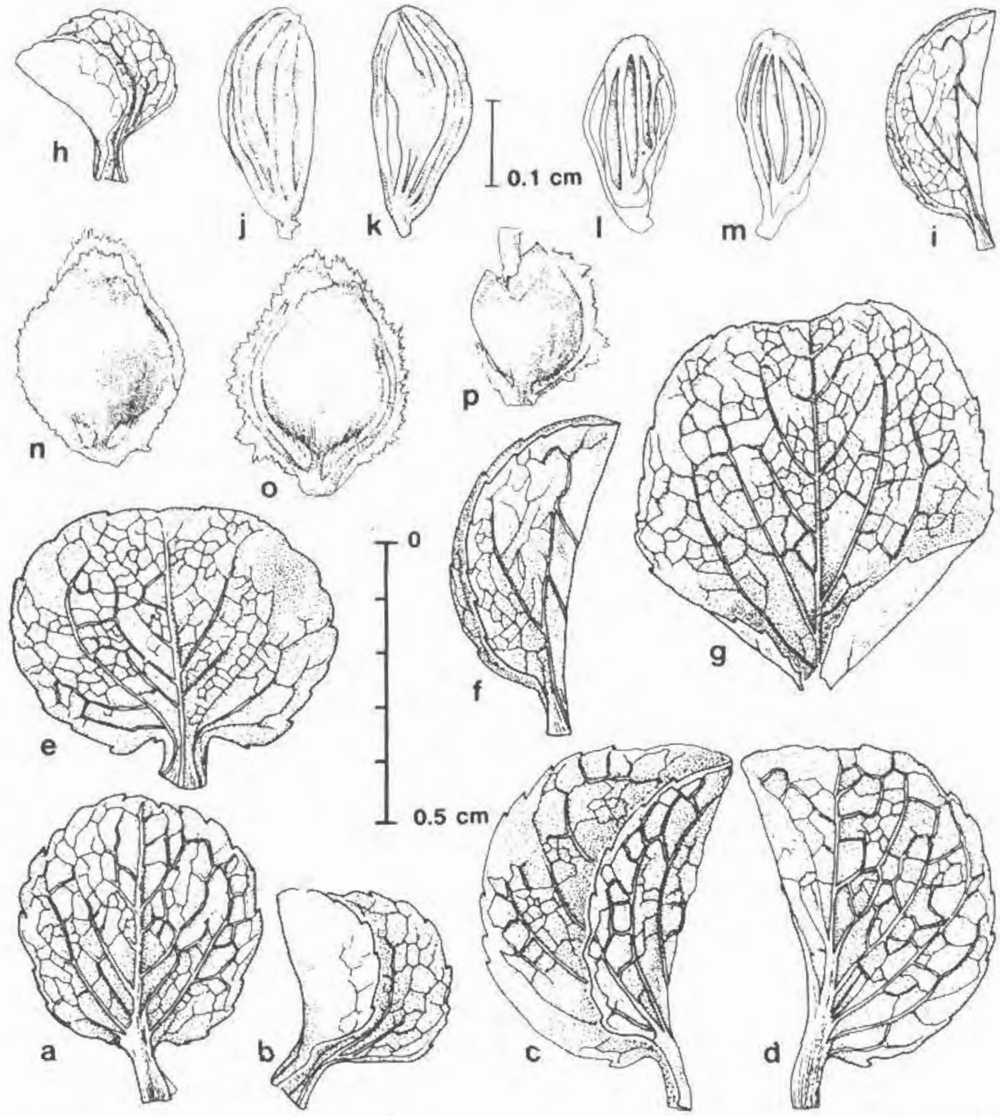

Figure 7. Selected plant fossils from Matianuck Ave. site, Windsor, CT. a-i, leaves of *Salix herbacea*, note that some of the leaves are folded in half, which is characteristic of fossils of this species; **j-m**, achenes (fruits) of *Thalictrum alpinum*, **j** and **k**, **l** and **m**, show both sides of the same fruit; **n-p**, achenes (fruits) of *Oxyria digna*, marginal wing eroded to different degrees. (**a-i**, lower scale; **j-p**, upper scale).

Conditions at and following lake drainage

At the end of its existence Lake Hitchcock was very shallow in the southern part of the Valley; the lake bottom was only 10-12 m below lake level. North of the Holyoke Range in Massachusetts, the lake bottom was much deeper, 30-50 m below stable lake level (see fig. 4 in Koteff and others, 1988). When the the relatively narrow, easily erodable was breached, water levels must have dropped relatively quickly. Once the lake level had lowered by 10-12 m, however, the lake-bottom surface was exposed, the lake in the south was gone, and the incipient Connecticut River began the more difficult task of entrenching the lake bed over a very long distance. Extensive fluvial terrace deposits at several levels (the highest is seen at Stop 5) record this entrenchment of the lake-bottom surface. At this time the lakebed in the southern part of the basin was acting as the dam for a lower-level remnant-lake that occupied the northern part of the basin. Water in this remnant lake would still have been 20-40 m deep. The Ridge and Larsen (1990) estimate of lake drainage at 12.4 ka at the Canoe Brook site in southern Vermont most likely applies to this northern remnant of Lake Hitchcock.

Evidence for an open landscape and strong northwesterly winds after the lake drained is preserved on the lakebed in the form of extensive longitudinal and transverse dunes especially on the east side of the Valley. Dunes up to 9-m high can be seen along the route between Stop 5 and 6 (fig. 11); many more are present on surfaces between Stops 5 and 7. The dune exposed at Stop 7 is on a high lake-bottom surface at 175-185 ft in altitude in front of a delta graded to a high level of glacial Lake Hitchcock (20 ft (6 m) above stable level); the stable shoreline at that locality was at 170 ft. The north to northeasterly wind directions indicated by the dune morphology and internal structure (see discussion by R.M. Thorson, Stop 7) record an early wind direction after the lake had dropped to stable level, and while the ice margin was still in southern Massachusetts. The upper section of this dune contained detrital pieces of charcoal which yielded a ^{14}C date of **11,485±115 yr BP** (AA-7154), (L. McWeeney, pers. comm.); thus, vigorous eolian activity continued at least until this time.

The incision of the lakebed by the early Connectiuct River down to the level of the modern floodplain was probably accomplished by about 8 ka. We have recently obtained vibracores from two low-terrace areas in Massachusetts, 2-3 m above the modern Connecticut River floodplain. Basal organics in a core from the Huntington Road site (Hadley, MA) yielded a date of **9370±100 yr BP** (Beta-52257) and 35 cm higher in the core, **8780±80 yr BP** (Beta-52256); basal orgainics in the Sanderson core near Stop 8 (Whately, MA) yielded a date of **8530±110 yr BP** (Beta-52258) (L. McWeeney, pers. comm.).

CONCLUSIONS

Interpretations of the conditions under which the last ice sheet receded during the late Pleistocene vary with the evidence used, location of the study and the investigator. In southern New England the glacial sediment record indicates fairly rapid recession following the maximum extent of the Laurentide ice at about 21 ka (Sirkin, 1982; Stone and Borns, 1986). Deglaciation was characterized by systematic northward retreat of an active-ice margin fringed with a zone of stagnant ice and by morphosequence deposition (Koteff and Pessl, 1981; Stone and others, 1992; Stone and Schafer, in prep). In southern New England, ice retreat was dominated by meltwater (Gustavson and Boothroyd, 1987). By 12 ka, the ice margin was in southern Canada and marine waters occupied the St. Lawrence lowland (Dyke and Prest, 1987). The calculated retreat rate for southern New England is 50-75 m/yr, and the conventional wisdom of glacial geologists has been that rapid meltwater-dominated deglaciation occurred in a rapidly warming climatic regime. However, if we examine the paleobotanical literature for southern New England, we find that pollen records and plant macrofossils indicate the presence of cold-climate, tundra vegetation for several thousand years following deglaciation (see recent regional summaries by Gaudreau and Webb (1985) and Jacobson and others (1987)). Evidence for Connecticut comes from records at Rogers Lake (Davis and Deevey, 1964; Davis, 1965; 1969), Totoket Bog and Durham Meadows (Leopold, 1955; 1956), and Cedar Swamp, (Thorson and Webb, 1991). These records all indicate that between about 15 ka and 12.5 ka a treeless plant assemblage was present in southern New England characterized by sedges, shrubs and grasses (the "herb zone"). This assemblage is indicative of arctic to subarctic conditions.

Northward retreat of the ice margin from the Long Island terminal moraines through southern New England took place early (19-14 ka), before the great bulk of the Laurentide Ice Sheet began to dissipate. Retreat through southern New England was probably induced by changes in dynamic equilibrium within the ice sheet rather that climatic amelioration. A periglacial climate which supported permafrost, at least in some areas favorable to its development, persisted for several thousand years after the ice had left the area.

FIELDTRIP STOPS

STOP 1: HAIN BROTHERS SAND AND GRAVEL PIT, town of Windham, CT, Willimantic quadrangle. From Rt. 32 and Rt. 203 junction, proceed 0.75 mi (1.2 km) northeast on Rt. 203. Entrance to pit is left off Rt. 203, 0.75 mi northeast of Rt. 32 intersection. Pit is areally extensive; we will examine sections at nine exposures lettered **a-i** in fig. 7.

We will spend a couple of hours in this pit, and as we examine and perhaps argue about these intriguing "wedge structures", let us remember fondly Phil Schafer who spent much time here over the years, who first interpreted them as ice-wedge casts, and from whom we gained a great deal of our knowledge about the details of these structures.

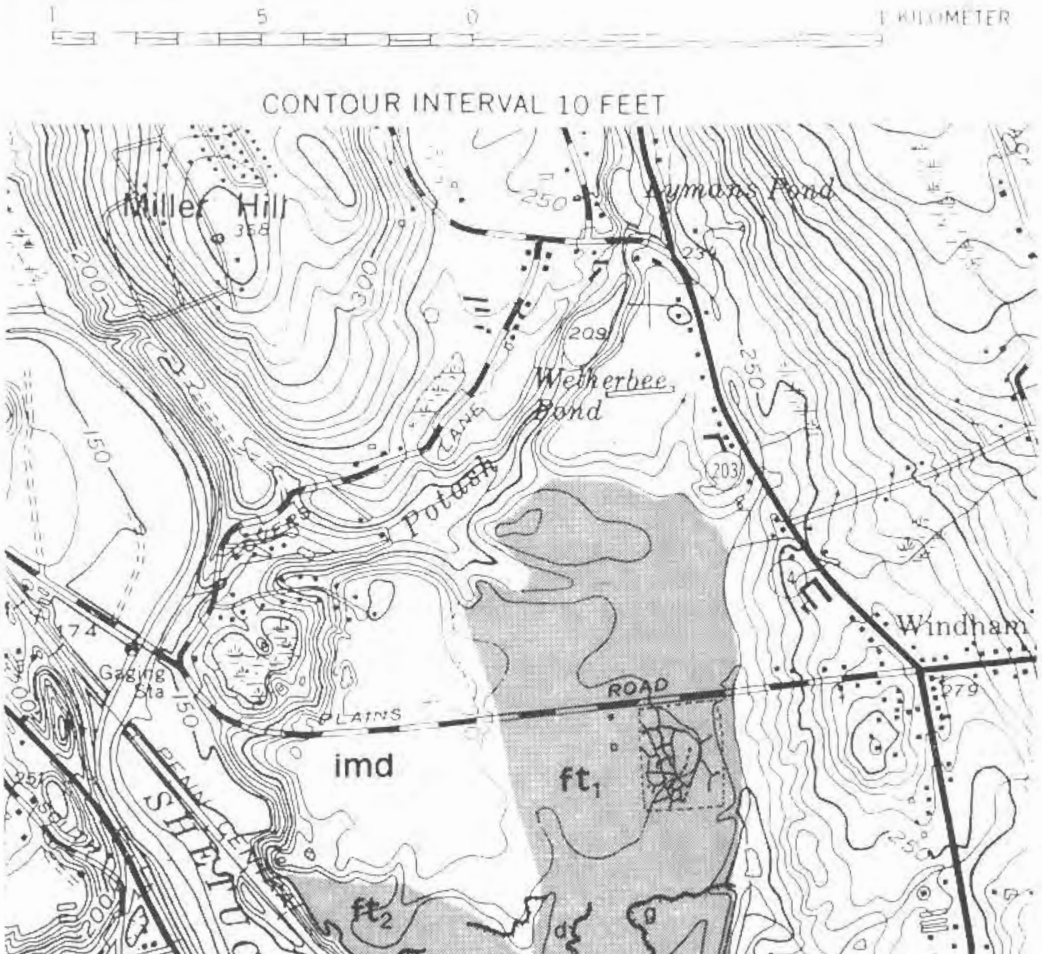

Figure 8. Topography of Hain Pit area. Pit excavation scarps (1990) shown by hachured lines. Polygonal ground pattern sketched from airphoto (fig.4). Shaded areas are fluvial terraces (ft_1, ft_2); these are inset into an ice-marginal delta (**imd**). Sections **a** through **i** described below.

The glacial geology in the vicinity of the Hain pit was mapped at 1:24,000-scale by Clebnik, (1980), who also recognized the occurrence of numerous wedge-shaped structures in the Hain pit. More detailed analysis of the area during investigation of the wedge structures, has shown that the area included in figure 8 consists of three different episodes of meltwater deposition. An ice-marginal delta (**imd**) was deposited into a small drift- dammed lake in the Shetucket valley (one of a series of such lakes in that valley). This delta has an ice-marginal slope on its NW side and a surface plain that slopes southeast from 255 ft altitude near the ice-margin position to 235 ft near its southern edge where it is cut by a 20-ft scarp; its eastern side is cut by a 10-15-ft scarp. Both scarps were incised during the succceeding episodes of meltwater erosion and deposition. The 245-250-ft surface SE of Wetherbee Pond is also part of the ice-marginal deltaic sequence; there are no exposures in the delta surface, but it is estimated that the delta was built into a lake with a water plane of about 230 ft altitude.

The second episode of meltwater deposition was initiated when lake levels in the Shetucket valley dropped (probably due to erosion of the drift dams) by at least 25 ft. This resulted in incision and reworking of the surface of the ice-marginal delta by meltwater that flowed down the Potash Brook valley east of Miller Hill from an ice margin position farther to the north. The surface of this fluvial terrace deposit (**ft**$_1$) slopes from 225 ft at the north end, through the patterned ground field (fig. 4), and the Hain pit area to about 215 ft where it is cut by a 10-ft scarp visible only on older (pre-gravel pit) airphotos. The deposits of this fluvial terrace are exposed in sections **d** through **h** in the Hain pit; they comprise ~3 m of horizontally layered and cross-bedded sand and pebble gravel. This material is the main source of gravel in the Hain pit, and most of the wedge structures deform these beds. Near the lower part of these sections, a disconformable contact between the fluvial sands and gravels and the underlying delta foreset and bottomset beds of the deltaic sequence (**imd**) is exposed in some places.

The third episode of meltwater deposition took place slightly later, perhaps at first contemporaneously with the **ft**$_1$ terracing. Distal meltwater in the Shetucket valley terraced the southern edge of the ice-marginal deltaic deposits and constructed a 205-ft terrace surface (**ft**$_2$). Fluvial sand and pebble gravel beds exposed in section **i** are as much as 6 m thick.

Section a: reveals approximately 12 m of sandy delta forset beds; the altitude of the pre-excavation stream-terrace surface here was 195 ft; only a small remnant of the original surface remains; the floor of the excavation is now at an altitude of about 150 ft.

Section b: reveals sandy and silty lacustrine beds higher in section than section **a;** quite spectacular penecontemporaneous water-escape structures can be seen near the top of this section, which is not the original surface. Several meters of section have been stripped by excavation.

Section c: reveals about 5 m of sandy and silty lacustrine beds, still higher in section than previous section **b**; well-developed sets of climbing ripples can be seen here, and also more water-escape/load deformation in the form of flame, and ball-and-pillow structures. Periodically exposed near the top of this section is a 0.3 to 1.0-m-thick and 15-m-long lense of varved silt and clay in 10 couplets (each about 5-10 cm thick). These fine-grained beds result in a perched water table on the floor of much of the pit area to the north. About 3 m of fluvial sand and gravel beds have been removed from the top of this section; the original terrace surface was at 205 ft in altitude.

Sections **d** through **i** are cut into the original ground surface; wedge structures are exposed (or formerly were) in all of these sections. All sections except **f** have been stripped of 0.5-1.0 m of upper soil horizon prior to sand and gravel extraction.

Section d: fluvial terrace deposits (ft$_1$) are exposed in this area. At times the disconformity between the fluvial beds and the underlying deltaic/lacustrine beds is exposed near the base of this section. Lacustrine material underlies the pit floor; puddles of water on the pit floor indicate the presence of a perched water table above the underlying fine-grained lacustrine beds. Several wedges which deform the fluvial beds and penetrate into the lacustrine section, have been visible along this N-S trending face during the past 2 years, as the face has moved eastward. The most prominent wedge, near the center of this face, strikes nearly E-W and is exposed in section **g**, as well; its linear ground trace across the topsoil-stripped surface between **d** and **g** can be seen on the 1990 airphotos.

Section e: exposes about 3 m of fluvial section (ft_1), beds are generally finer-grained than in section **d**, consisting mostly of fine to medium sand, with only minor pebble gravel. Three wedges, all of which exhibit preserved upturned beds and pressure-effects in the wall strata are exposed in the pit face (see fig 3. and discussion in text). One wedge here strikes N40°W and intersects with its neighbor, which strikes N60°E, approximately 6 m in front of the pit face. This intersection can be seen by scraping the pit floor and following the deeply oxidized sand trace of the narrow "at depth" wedge fillings.

Section f: is a ~1.5-m-deep backhoe excavation in the only remaining non-pitted, non-topsoil-stripped surface in the Hain pit; the trench is cut perpendicular to a wedge fracture. Here we can see an intact modern soil horizon developed in the eolian mantle. "Pebbles" within the eolian fine sand have been incorporated from the underlying fluvial gravel and sand beds by cryoturbation. Note also that the "soil" is not deformed by the wedge structure. The following is a pedon description outside of the wedge structure by Dr. Harvey Luce, University of Connecticut, Storrs, CT:

<u>Phase</u>: Agawam gravelly fine sandy loam, 0-3% slope.

Ap- 0 to 23 cm; very dark brown (7.5YR2/2) gravelly fine sandy loam; weak fine subangular blocky structure parting to weak fine granular structure; very friable; abundant roots; abrupt smooth boundary.

Bw1- 23 to 37 cm; dark yellowish brown (10YR4/6) gravelly fine sandy loam; weak fine subangular blocky structure; very friable; many roots; gradual smooth boundary.

Bw2- 37 to 51 cm; yellowish brown (10YR5/6) fine sandy loam; weak medium subangular blocky structure; very friable; common roots; abrupt smooth boundary.

2C- 51 to 101 cm; brownish yellow (10YR6/6) gravelly sand; single grain; loose; very few roots.

Section g: Wedge intersections exposed in this face in the past have displayed extremely complex deformation structure. Presently exposed is a deep E-W striking wedge into which the eolian plug material penetrates as much as 3 m. As a result of the greater depth at which the eolian material lies in many of the wedges but particularly in this one, it is gray in color. This is because the deeper eolian material is not pervasively oxidized by B-soil horizon processes as it is in the shallower eolian mantle material.

Section h: This section has historically exposed numerous soil-involution (cryoturbation) structures, some of which are developed in the top of the wedge structures. The involutions appear to be of type 2 and 4 of Vandenberghe (1988), formed by loading due to cryohydrostatic pressures. These kinds of involutions are generated during the progressive degradation of permafrost, when a lot of water is present but an underlying impermeable horizon (the permafrost) is still present. As discussed by Van Vliet-Lanoe (1988), the presense of fine-grained eolian cover on coarser-grained fluvial materials, produces a positive frost susceptibility gradient. With progressive lowering of the water table associated with the downward migration of permafrost table, pot-shaped involutions form in ice-wedge cracks, especially at their intersections which are the lowest and wettest points of the microtopography.

Section i: For the past several years, this section has revealed about 6 m of fluvial pebble gravel and sand (ft_2); no wedges have been seen here during this time, however their presence here is indicated by Thorson and others (1986). At present this excavation has expanded northward by about 150 m from the 1990 scarp shown on figure 8. Excavation has begun to intersect the south end of unit **imd**.

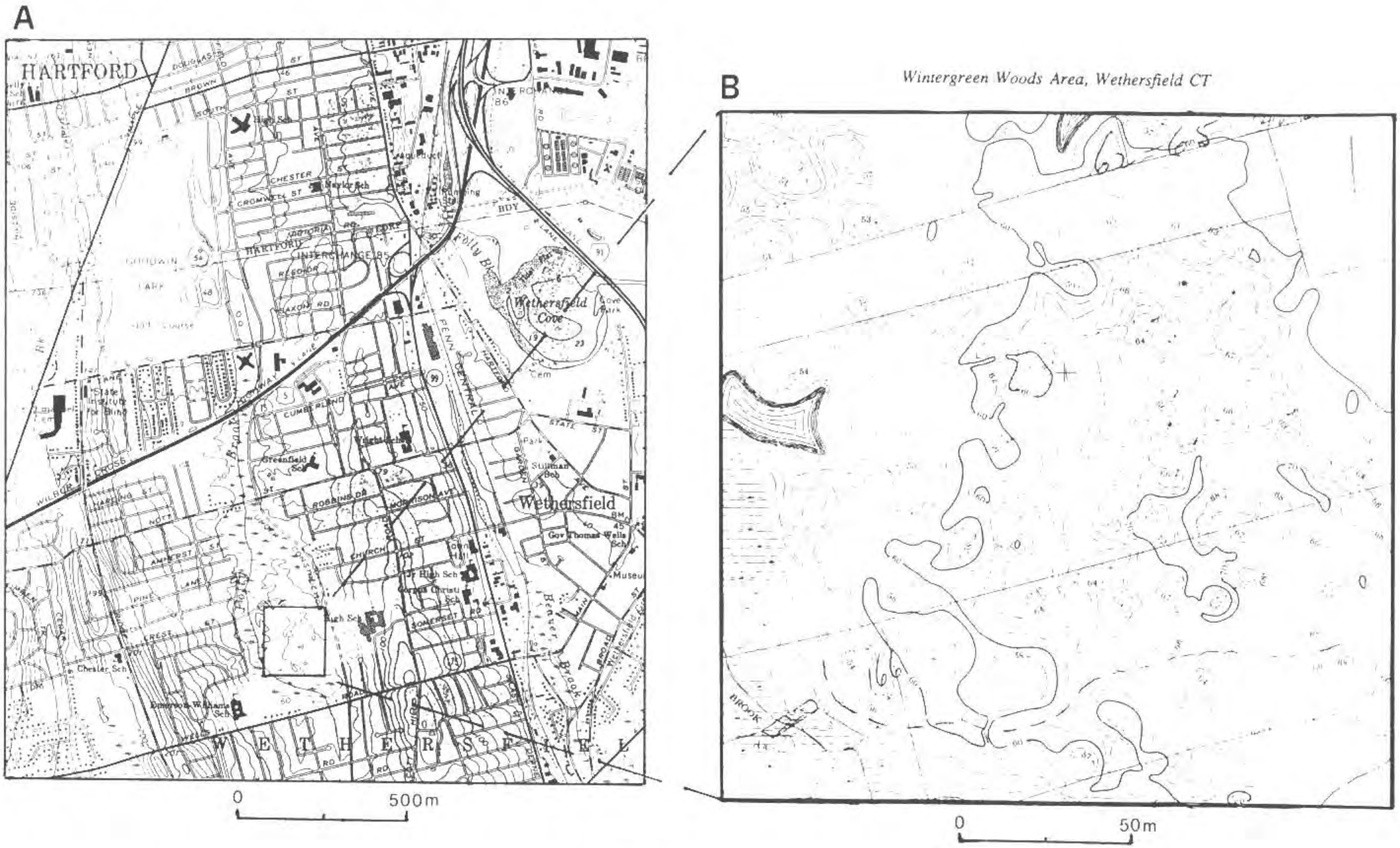

Figure 9. Topography of Wintergreen Woods Area. **A)** Section of Hartford South 1:24,000-scale quadrangle map with 10-ft contour interval (reduced). **B)** Section of Metropolitan District Commission (MDC) 1:2400-scale map 147 with 2-ft contour interval (reduced).

STOP 2. WINTERGREEN WOODS park, town of Wethersfield, CT, Hartford South Quadrangle. From its junction with Silas Deane Highway (Rt. 99), proceed 0.3 mi (0.5 km) west on Wells Road (Rt. 175). Turn right on Folly Brook Drive. Entrance to park is at the end of Folly Brook Drive, 0.15 mi (0.25 km) north of Wells Road.

This is the best known locality to see the landforms which we have interpreted as pingo scars; park nature trails wind around the rims of the many depressions that occur on this surface. The lake-bottom surface here is part of a finger of Lake Hitchcock that existed in a low area between two till/bedrock hills. Although it is not at all apparent on the quadrangle map (fig. 8A), the MDC map (fig. 8B) with a 2-ft contour interval shows that the entire surface is covered with shallow closed depressions and depressions with breached rims marked by C-shaped contours. As is generally the case, the water table is at relatively shallow depth beneath these lake-bottom surfaces, and its seasonal fluctuations result in alternately ponded (or marshy) and dry conditions in the lower part of the depressions. Many of them are in fact "vernal pools" which are homes to numerous amphibious creatures (some of which are endangered species). The pingo-scar depressions in this area are relatively small compared to other areas, ranging in diameter from 8-25 meters. In this area the features are present on surfaces between 50 and 70 ft altitude which is as high as the lake-bottom surface gets here; the stable level shoreline was at an altitude of 82 ft. The features are conspicuously absent on the lower surfaces incised into the lake-bottom by the north-draining Folly Brook. This general characteristic of their distribution demonstrates their early-postlake time of formation.

STOP 3. I-91 CUT NORTH OF HARTFORD, town of Windsor, CT, Hartford North Quadrangle. From Windsor Ave. (Rt. 159) at Wilson, proceed west on Matianuck Ave; 0.4 mi (0.6 km) north of Rt. I-91 underpass, turn right on Oak St. (cul de sac). Roadcut was formerly behind highway fence; neighboring pingo scar is in yard on right.

A 4-m-deep, 165-m-long roadcut along the west side of southbound Rt. I-91 between exits 34 and 35 was exposed during the summer of 1989. The cut is now graded over; we include it as a STOP because of its importance to the interpretation of the rimmed depressions as pingo scars; this road cut remains the only cross-sectional exposure we have seen. The section (fig. 5A) revealed upper Lake Hitchcock bottom sediments, about 3 m of rhythmically bedded vf-f sand and silt containing pervasive water-escape structures mostly in the upper 1-2 m overlying poorly exposed varved silt and clay. The section cut through a rimmed depression (fig. 5B) developed in the lake-bottom sediments. Topographically, the feature was a circular, 40-m-diameter, 3.5-m-deep wetland depression (fig. 5B) similar in morphology to many other features in the area. The cross section clearly showed that the surface depression was underlain by deformed lake sediments. The parallel-bedded lacustrine sediments could be traced toward the depression where they were upwarped beneath the surface rim, then intensely collapsed downward beneath the central part of the depression. The structural depression created by the collapsed beds was partially filled with a bowl-shaped body of eolian-supplied fine to medium sand overlain by a body of peat and gyttja (see previous discussion of internal structure and ^{14}C dates above). A feature we interpret as a small ice-wedge cast cut lacustrine sediments at the south end of the exposure (see fig. 5A).

Connecticut Department of Transportation test borings indicate that 15-18 m of varved silt and clay underlie this section. The presence of deformed lake deposits below the surface depression exposed in this roadcut eliminates the alternative hypothesis that these features are eolian deflation hollows.

STOP 4. MATIANUCK AVE. ROADCUT, town of Windsor, CT, Hartford North Quadrangle. From Oak St., proceed 0.4 mi (0.6 km) north on Matianuck Ave. to road construction for new entrance ramp for relocated Rt. 291 (Bissell Bridge connector). East-west trending roadcut passes beneath Matianuck Ave. Section on east side was still visible in August 1992.

This roadcut is slowly being graded over and may not be viewable in the near future. The lake-bottom section exposed here is similar to that formerly exposed at the I-91 cut, Stop 3. West of Matianuck Ave., the varved section was better exposed than anywhere at the I-91 cut. A composite sketch of sections east and west of Matianuck Ave. as seen in Oct. 1991 and March 1992, respectively, is shown in figure 10.

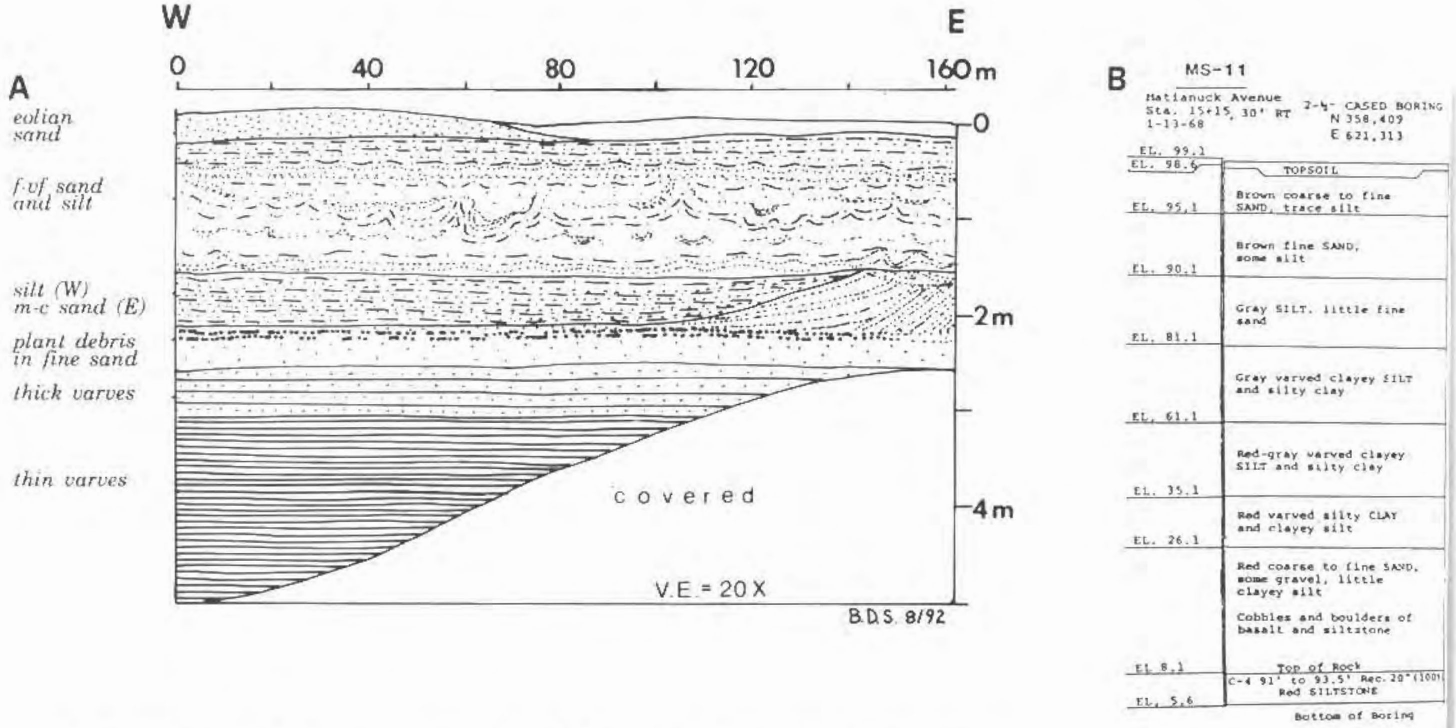

Figure 10. A) Composite sketch of Matianuck Avenue roadcut section **B)** Bridge test boring log for Matianuck site.

West of Matianuck Ave, the deeper section exposed 2.5 m of varved silt and clay. Two meters of thin gray varves averaging 1-cm in thickness were overlain by 0.5 m of thicker reddish gray varves averaging 2-3 cm in thickness. Test borings for bridge construction (fig. 10B) indicate the presence of approximately 20 m of varved silt and clay at depth below this section. The varved section recorded in boring logs shows typically red varves in the lower part, alternating red and gray varves in the middle of the section and gray varves in upper part. This color/mineralogic change within the varve section reflects the early presence of ice in the basin supplying local "red" (Mesozoic rock-derived) material, then changing with increasing distance of ice source to "gray" (crystalline rock-derived) material supplied by streams flowing from the highlands into the lake.

Approximately 2.5 m of fine sand and silt beds containing pervasive water-escape structures overlies the varved section. At the eastern end of the roadcut, a lense of medium to coarse sand in a 1.0-m-thick set of crossbeds (containing abundant large muscovite flakes) prograded westward intertonguing with thin (0.5-1.0 cm-thick) parallel-laminated fine sand beds. Two of these beds consisted predominantly of detrital plant debris. The organic material yielded a ^{14}C date of **14,100±110 yr BP**. The well-preserved state of plant macrofossils indicates that the plant debris had not been transported far. Local and regional stratigraphy indicates that the sandy beds containing the plant material were deposited in association with lowering of the lake level by breaching of the drift dam (see previous discussion).

STOP 5. EAST HARTFORD SAND AND GRAVEL PIT and KELSEY-FERGUSON CLAY PIT; town of South Windsor, CT, Manchester quadrangle. From junction of Rt. 291 (Bissell Bridge) with Rt. 5, proceed 2.5 mi (4 km) north on Rt. 5. Turn right at light onto Strong Rd., then immediate left. Turn right past the K-F Brick Co. buildings onto dirt road. Two exposures are present; the sand pit exposes stream terrace deposits and a deeper clay pit exposes lake-bottom sediments.

As seen in figure 11, stream terrace deposits (**Qst**), 3-4 m thick, overlie lake-bottom deposits (**Ql**) which are as much as 53-m thick in the vicinity of this Stop. The stream terrace deposits are pebbly, crossbedded sands laid down soon after the draining of Lake Hitchcock. The terrace surface is at 75-85 ft altitude here and grades to a 50-ft terrace surface cutting through the drift dam deposits to the south. Its surface has been modified by eolian activity. The terrace deposits are inset only slightly into the underlying lake-bottom surface which lies at 65-70 ft altitude beneath it. Lake-bottom surfaces just to the east (Stop 6) are at 85 ft altitude. Stable lake level here was at 115 ft altitude; after lowering of only 40 ft (12 m), the lake-bottom was exposed and terracing began.

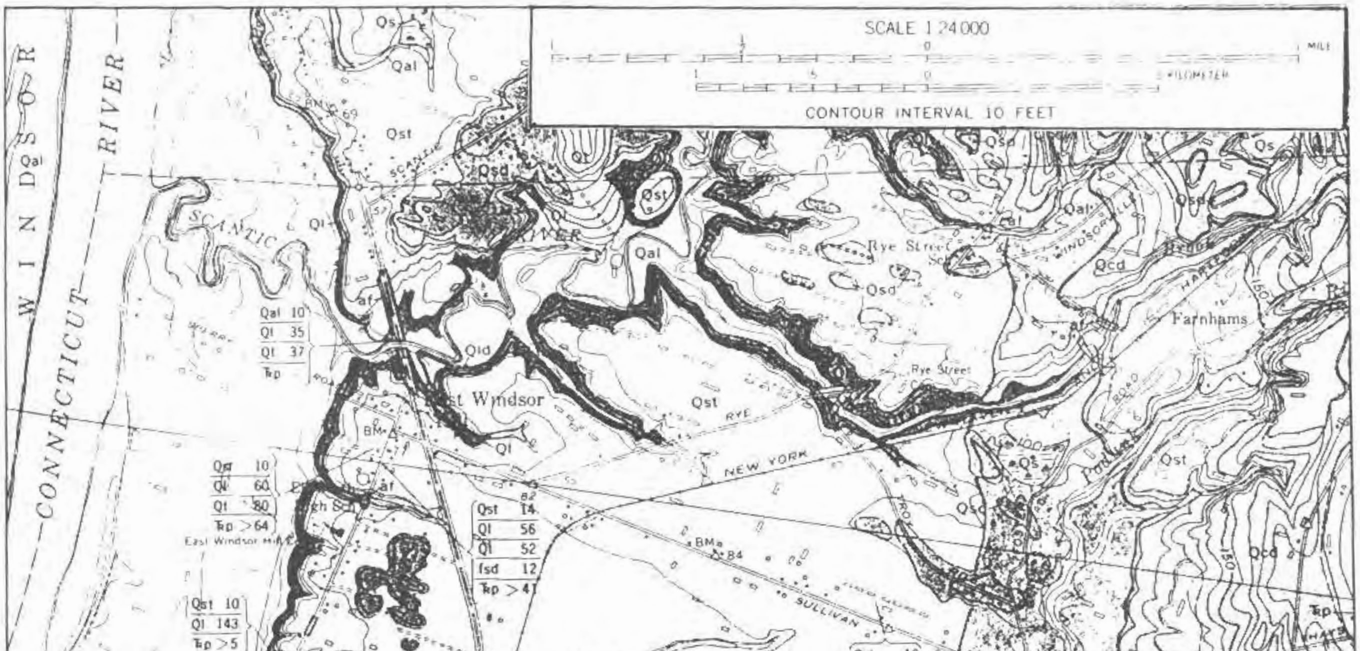

Figure 11. Topography/Geology in the vicinity of Stops 5 and 6. Modified from the geologic map of the Manchester quadrangle (Colton, 1965). Note that the contact between Qst and Ql has been changed. Qst-stream terrace deposits, Ql-lake-bottom deposits of glacial Lake Hitchcock, Qsd-eolian sand dune deposits, Qs-swamp deposits, Qal-floodplain alluvium. Area of Figure 12 is outlined.

About 4 m of varved sediments are periodically well exposed in the clay pit. This location is 2.5 km from the lake margin and is far removed from major deltaic sources of sediment. The lake at the time of deposition was 17-21 m deep. Sediments are rhythmically bedded dark clays and light colored multiple-graded silts. However, there is enough inconsistency in the bedding (i.e. thin clay bands in silt and thin silt layers within the clay) to present problems in interpreting these rhythmites as varves. It is difficult to determine which silt laminae represent the commencement of a yearly deposit and which are the result of a chance introduction of silt into the lake during clay deposition. It is likely that sediments were transported to the site by overflows and/or interflows.

EN ROUTE TO STOP 6: Longitudinal and transverse dunes up to 9 m high built by northwesterly winds occur on the terrace surface. Strong Road crosses several dunes and Foster Road follows the crest of one. The wetland areas (Qs) on the terrace surface are relatively large; these are deflation areas associated with the dunes. Near its junction with Foster Road, Strong Road crosses onto a nonterraced lake-bottom surface (hand shaded and label Ql on figure 11) that lies at 85-95 ft altitude. This surface has many small, shallow closed depression/wetlands and breached, "C"-shaped depressions (visible on the 2-ft contour interval map in figure 12), but not visible on the 10-ft contour interval map of figure 11. Small sand dunes may be present on this surface as well, but it is difficult to distinguish them from the pingo-scar rims.

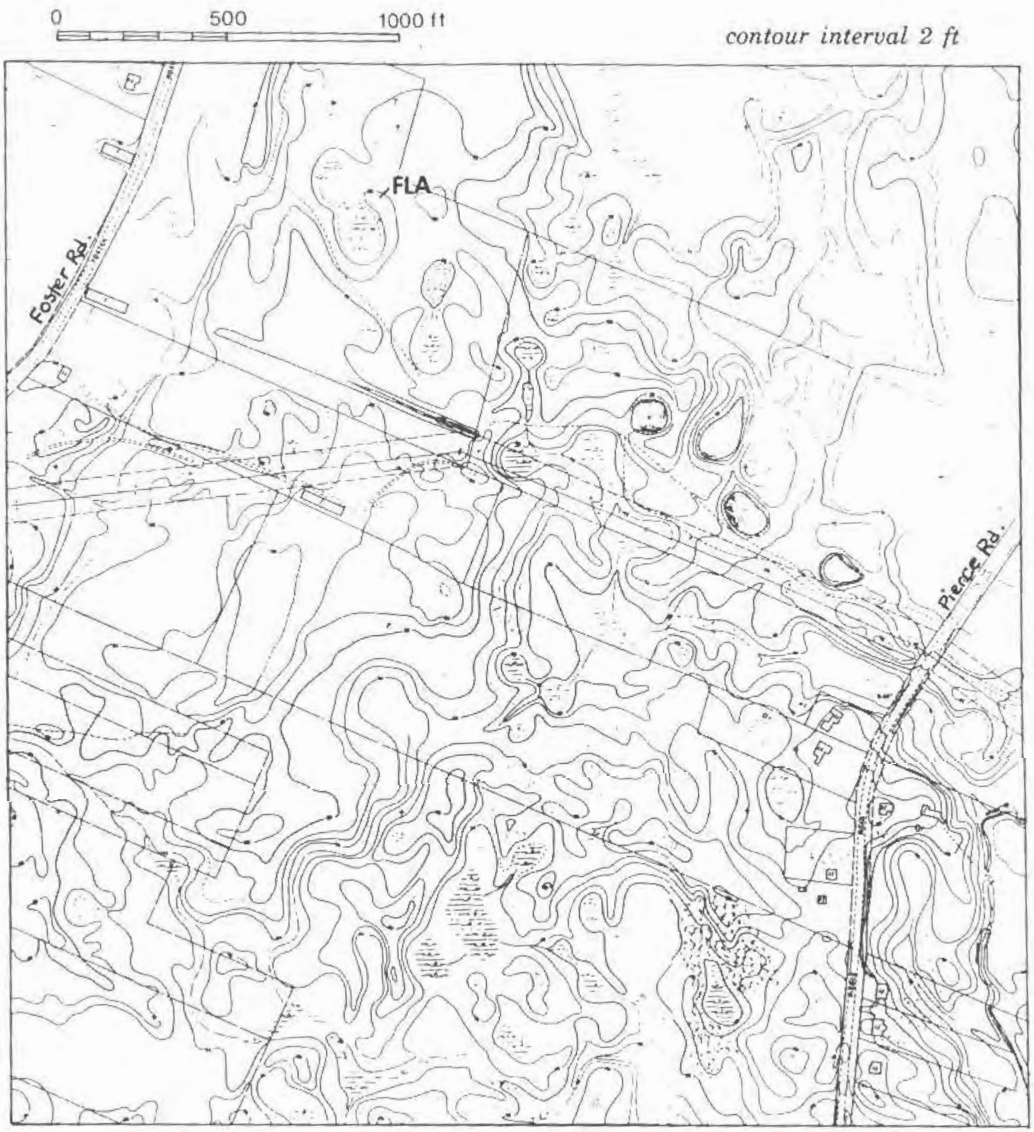

Figure 12. Topography in the vicinity of Stop 6. Section of MDC map 467 showing detailed, large-scale topography of Wapping area pingo scars. See figure 11 for regional location. Path along power lines between Pierce Road and Foster Road provides access to view some of the best-developed of these features.

STOP 6: WAPPING PINGO SCARS: town of South Windsor (village of Wapping), CT, Manchester quadrangle. Proceed east on Strong Road, turn right on Foster Road. Proceed 0.25 mi (0.4 km) to Seventh Day Adventist Church on left. Trail begins at small outbuilding at edge of parking lot.

The trail between the church parking lot and Pierce Road to the east (see figure 12) follows a line of well-developed pingo scars. These features are larger on avèrage than the ones at Stop 2. These are generally in the 30- 80-m diameter range. The ones to the east along the power line are deeper and contain perenially ponded water (they show up as small ponds on the 1:24K quadrangle map, fig. 11). We have obtained a number of vibracores from several of the depression nearest to the church parking lot; cores and GPR lines from pingo-scar **FLA** (fig. 12), are described in Oakes (1992).

STOP 7: ENFIELD DUNE: town of Enfield, CT, Springfield South quadrangle. Procced east on Rt. 220 (Elm St.) which becomes Shaker Road. Turn left on Washington Road; proceed 1.6 mi (2 km) and turn right on Brainard Road; dune lies along south side of Brainard Road; exposure is behind apartment complex now occupying a former sand pit where internal structure was once well-exposed.

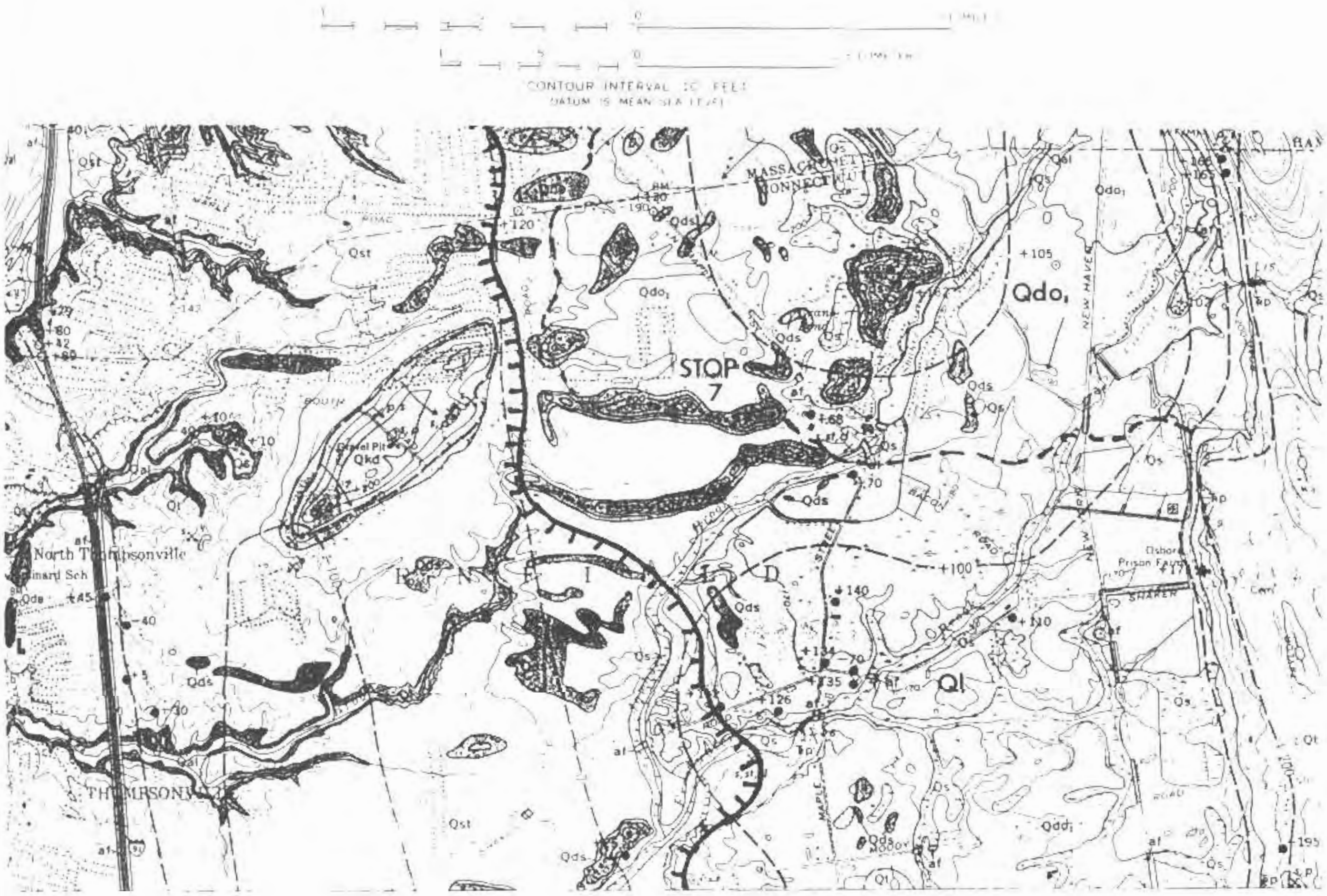

Figure 13. Topography/Geology in the vicinity of Stop 7. Modified from the geologic map of the Springfield South quadrangle (Hartshorn and Koteff, 1967). Lake Hitchcock stable-level shoreline is shown by heavy ticked line. Delta labeled Qdo_1 records a 190-ft lake-level, 20 ft (6.5 m) higher than stable level. The 175-185-ft surface to the south of the delta is a lake-bottom surface (Ql) associated with the higher lake level. The east-west trending transverse dunes lie on this surface; the high lake-bottom surface SE of the dunes displays many pingo scars (not visible at the scale of this map).

Description by Robert M. Thorson, University of Connecticut, Storrs, CT:

The "Enfield dune" is a complex elongate stationary dune that extends for nearly 1 km in an east-west alignment. The external form of the dune is irregular, but in its central section, the south-facing slope is consistently steeper than the northern one. This asymmetrical topographic profile suggests that it is a transverse dune, formed by northerly winds.

The internal structure of the dune was documented by Schile (1991). The bulk of the dune consists of parallel, inversely graded laminae that dip gently to the north-northeast, and which are interpreted as subcritical translatent strata formed by southeasterly migrating ripples on the northerly stoss side of the dune. The southerly part of the dune is dominated by grainfall and grainflow strata in steeply dipping beds, which formed on the lee side of the dune below its crest. Secondary structures are generally limited to burrow casts and root mottles near the present dune surface.

The so-called eolian mantle caps the dune with 1-2 m of silty unbedded sand, the bulk texture of which is similar to the underlying well-bedded material. Lying conformably below the eolian mantle and unconformably above the stratified sand is a complex transition zone ranging from a few centimeters to a meter thick. The base of the transition zone typically exhibits one or more horizons in which coarse sand and granules are intermixed with the sand. The sand above the granule concentration(s) is weakly and discontinuously bedded, but rarely contains shallow trough cross-stratified lenses of coarser sand interpreted as the fill of surface rills. At the crest of the dune, the transition zone contains an incipient soil (expressed as an oxidized, mottled, and bioturbated horizon) that is unconformably overlain by crosscutting beds of eolian sand 5-20 cm thick with trough-shaped basal contacts. One of these beds contained fragments of detrital charcoal collected by Schile. This charcoal was indentified by Lucinda McWeeney as *Pinus sp.* and yielded a date of **11,485±115 yr BP** (AA-7154).

We interpret the stratigraphy of the Enfield dune to indicate the progressive growth of a stationary transverse dune. This took place in front of a delta on a high lake-bottom surface after lowering of the lake to stable level. The north-northeasterly wind regime is consistent with eolian erosional surfaces reviewed by Schafer and Hartshorn (1965), and requires the continued presence of a strong anticyclonic circulation set up by the Laurentide Ice Sheet. The stratigraphy, sediment texture and sedimentary structures of the transition zone are consistent with a gradual stabilization of the area by vegetation. A discontinuous cover of plants in the interdunal environment probably restricted the supply of sand prior to a reduction in wind intensity, causing erosion of the dune surface and a lag concentrate, which in turn, would have led to edaphic conditions more favorable for plant growth on the dune. The remainder of the transition zone suggests the continued saltation of sand over a progressively vegetated surface. Local erosion and re-activation of sand accumulation, possibly by winds with a more westerly component, is suggested by the dated horizon.

STOP 8: SANDERSON PIT: town of Whately, Mass., Mt. Toby Quadrangle. Proceed north into Massachusetts on Rt. I-91. Pass through the Holyoke Range. Exit Rt. I-91 at interchange 24; turn right at end of ramp; proceed 0.1 mi (0.16 km), turn right (east) on Rt.116, proceed 0.15 mi (0.24 km), turn right on Long Plain Road; proceed 0.4 mi (0.65 km) to Sanderson farm nursery. Pit is down farm road on left at edge of terrace scarp.

A small pit cut into the upper lake-bottom section near its edge bounded by the Connecticut-River-incised scarps is the focus of this stop. The cut reveals 3-4 m of rhythmically bedded fine sand and silt, similar in sedimentary structure to the upper lake sections at Stops 3 and 4 in the southern part of the basin. Well logs indicate the presence of varves below the exposed section; the exposed fine-sand and silt beds are deformed by pervasive water-escape convolutions. We will use this location to discuss the origin of these water-escape structures, often seen near the top of the lacustrine section. Were these structures generated by syndepositional loading?; are they cryoturbations generated by intense freezing or melting of permafrost?; or, were they generated by liquifaction induced by earthquakes associated with glacioisostatic rebound and lake drainage?

The lake-bottom surface (Qhl, figure 14) here has not been terraced; it lies at 185 ft altitude which is 120 ft (37 m) below the stable lake level recorded by the nearby Sunderland delta (topset/foreset contact at 295 ft). Stable Lake Hitchcock was much deeper here than it was in the south of the Holyoke Range. When the drift dam at the southern end of the lake was breached, waters in the south lowered only 10-12 m before intersecting the lake bottom. North of the Holyoke Range, a remnant lake, 20-m deep remained after the drift dam was breached. This slowly lowering remnant lake may have lasted for a significant period of time as the outlet stream incised the 80-km length of the lake-bottom surface to the south.

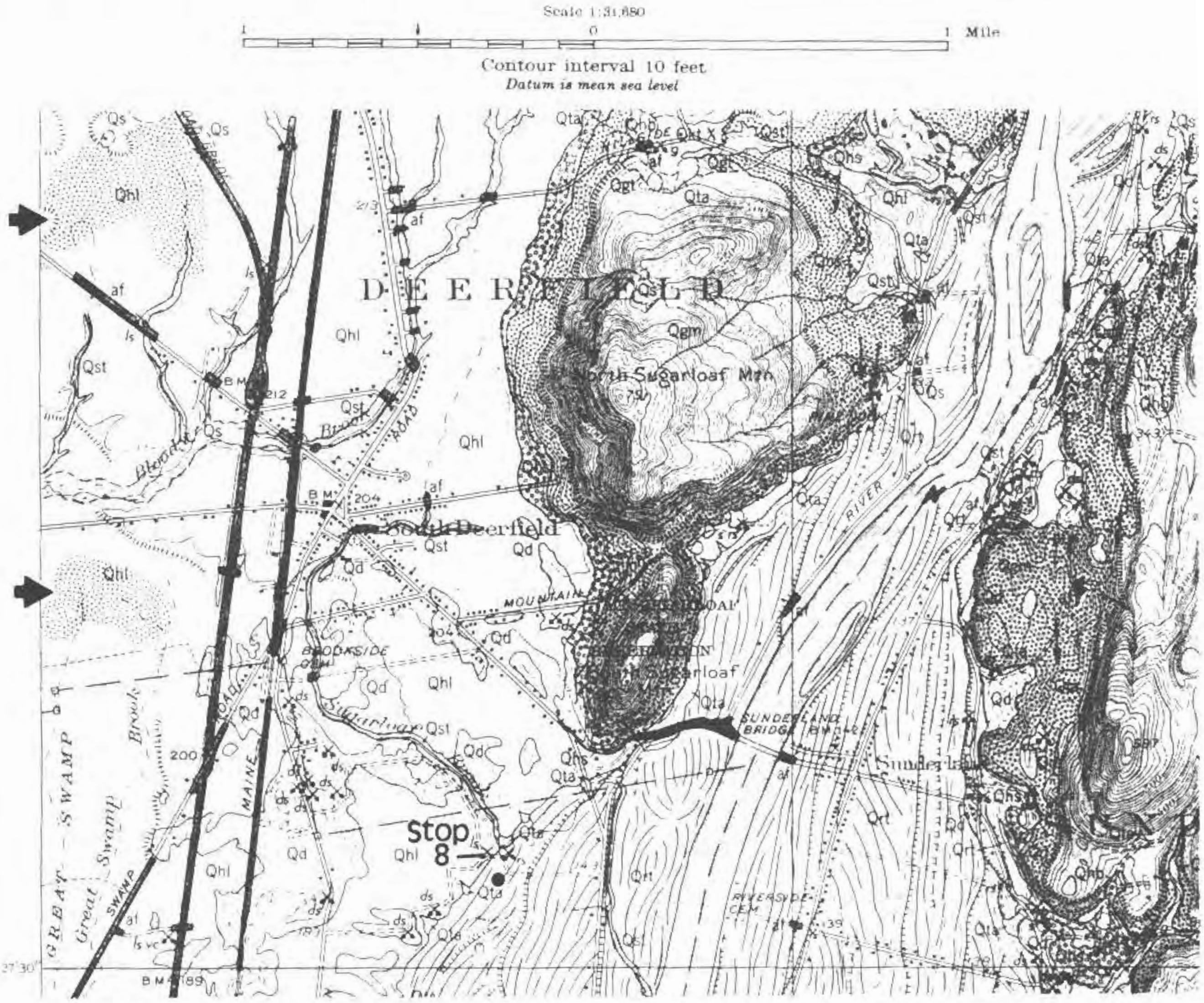

Figure 14. Topography/Geology in the vicinity of Stop 8. Section of the surficial geologic map of the Mt. Toby quadrangle, Massachusetts (Jahns, 1951). Arrows point to stippled areas mapped by Jahns as "kettled lake-bottom deposits"; these areas contain the small depressions we believe to be pingo scars. Dot is location of ^{14}C date on Connecticut River terrace.

The lake-bottom surface here also was subjected to intense eolian activity once the lake had drained. Dunes up to 6 m high built by northswesterly winds are common on this surface. Present also on this surface are the small rimmed depressions we believe to be pingo scars. These occur, in the area of figure 14 (see arrows), where Jahns (1951) shows a stippled pattern on unit Qhl to indicate "areas of kettled lake-bottom deposits".

A vibracore was taken on the 155-ft terrace surface (Qrt) below the pit; basal organic material lying on lake sediments yielded a ^{14}C dates (obtained by L. McWeeney) of **8530±110 yr BP** (Beta-52258); two other dates from basal organics on a similar low terrace on the east side of the river 11 km to the south were **9370±100 yr BP** (Beta-52257) and **8530±80 yr BP** (Beta-52256). Before 9 ka, the Connecticut River had incised down into the Lake Hitchcock lakebed nearly to the level of the modern floodplain.

REFERENCES

Black, R. F., 1974, Ice-wedge polygons of northern Alaska: in Coates, D.R., ed., Glacial Geomorphology, Fifth Annual Geomorphology Conference, Binghamton, N.Y. p. 247-275.

Black, R. F., 1976, Periglacial features indicative of permafrost: Quaternary Research, v. 6, p. 3-26.

Black, R. F., 1983, Pseudo-ice-wedge casts of Connecticut, northeastern United States: Quaternary Research, v. 20, p. 74-89.

Boothroyd, J. C. and Lawson, D. A., 1989, Wedge structures in southern Rhode Island: Evidence for postglacial permafrost conditions: Geological Society of America Abstracts with Program, v. 21, p.5.

Clebnik, S. M., 1980, Surficial geology of the Willimantic quadrangle, with map; Connecticut Geological and Natural History Survey, Quadrangle Report QR-39, 45 p.

Colton, R. B., 1960, Surficial geologic map of the Windsor Locks Quadrangle, Connecticut: U.S. Geological Survey Geologic Quadrangle Map GQ-137, scale 1:24,000.

Colton, R. B., 1965, Surficial geologic map of the Manchester Quadrangle, Connecticut, U. S. Geological Survey Geologic Quadrangle Map GQ-433, scale 1:24,000.

Davis, M. B. and Deevey, E. S., Jr., 1964, Pollen accumulation rates: Estimates from sediment of Rogers Lake; Science, v. 145, p. 1293-1295.

Davis, M. B., 1965, Phytogeography and palynology of northeastern United States, in Wright H.E. Jr., and Frey, D.G. eds., The Quaternary of the United States: Princeton, N.J., Princeton University Press, p. 377-401.

Davis, M. B., 1969, Climatic changes in southern Connecticut recorded by pollen deposition at Rogers Lake: Ecology, v. 50, no. 3, p. 409-422.

De Gans, W., 1988, Pingo scars and their identification, in Advances in Periglacial Geomorphology, New York, N.Y., John Wiley & Sons, Inc., p. 279-322.

Denny, C. S., 1936, Periglacial phenomena in southern Connecticut: American Journal of Science, v. 32, no. 191, p. 322-342.

Dyke, A. S. and Prest, V. K., 1987, Late Wisconsinan and Holocene history of the Laurentide Ice Sheet: Geographie Physique et Quaternaire 41(2): p. 237-263.

Flint, R. F., 1956, New radiocarbon dates and late-Pleistocene stratigraphy: America Journal of Science, 254: 265-287.

Gaudreau, D. C. and Webb, T. III, 1985, Late-Quaternary pollen stratigraphy and isochrome maps for the northeastern United States: in Pollen Records of Late-Quaternary North American Sediments, A.A.S.P. Foundation, p. 247-279.

Gustavson, T. C. and Boothroyd, J. C., 1987, A depositional model for outwash, sediment sources, and hydrologic characteristics, Malaspina Glacier, Alaska: A modern analog of the southeastern margin of the Laurentide Ice Sheet: Geological Society of America Bulletin, v. 99, p. 292-302.

Hartshorn, J. H. and Koteff, Carl, 1967, Geologic Map of the Springfield South quadrangle, Massachusetts-Connecticut; U.S. Geological Survey Geologic Quadrangle Map GQ-678, scale 1:24,000.

Hartshorn, J. H. and Schafer, J. P., 1965, INQUA Field Conference A, pt. 1, New England-New York: Lincoln, Nebraska, Nebraska Academy of Sciences, International Association of Quaternary Research, VIIth Congress, 92 p.

Jacobson, G. L., Jr., Webb, T. III, and Grimm, E. C., 1987, Patterns and rates of vegetation change during the deglaciation of eastern North America; in North America and adjacent oceans during the last deglaciation; Boulder, Colorado, Geological Society of America, The Geology of NOrth America v. K-3, p. 277-288.

Jahns, R. H., 1951, Surficial geologic map of the Mt. Toby quadrangle, Massachusetts; U.S. Geological Survey Geologic Quadrangle Map GQ-9, scale 1:31,680.

Johnson, W. H., 1990, Ice-wedge casts and relict patterned ground in central Illinois and their environmental significance: Quaternary Research, v. 33, p. 51-72.

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 archaeology, in Brigham-Grette, J. (ed.), Fieldtrip Guidebook AMQUA 1988, p. 169-208.

Koteff, Carl, and Pessl, Fred, Jr., 1981, Systematic ice retreat in New England: U.S. Geological Survey Professional Paper 1179, 20 p.

Larsen, F. D., 1979, Retreat of ice left eastern upland of Massachusetts subject to deep freeze; U.S. Geological Survey Professional Paper 1150, p. 58.

Leopold, E. B., 1955, Climate and vegetation changes during an interstadial period in southern New England: Yale University, New Haven, unpublished PhD thesis, 99 p.

Leopold, E. B., 1956, Two late-glacial deposits in southern Connecticut; Proceedings of the National Academy of Sciences, v. 42, no. 11, p. 863-867.

Mackay, J. R., 1979, Pingos of the Tuktoyaktuk Penninsula area, Northwest Territories: Geographie Physique et Quaternaire, v. 33, p. 3-61.

Mackay, J. R., 1988a, Pingo collapse and paleoclimate reconstruction; Canadian Journal of Earth Sciences, v. 25, p. 495-511.

Mackay, J. R., 1988b, The birth and growth of Porsild Pingo, Tuktoyaktuk Penninsula, District of Mackenzie: Arctic, v. 41, p. 267-274.

Mackay, J. R., 1990, Seasonal growth bands in pingo ice: Canadian Journal of Earth Sciences, v. 27, p. 1115-1125.

Miller, N. G. and Thompson, G. G., 1979, Boreal and western North American plants in the late Pleistocene of Vermont: Journal of the Arnold Arboretum of Harvard University, v. 60, p. 167-218.

Mitchell, G. F. 1973, Fossil pingos in Camaross Townland, Co. Wexford: Proceedings of the Royal Irish Academy v. 73, Sec. B, p. 269-282.

Oakes, L. A., 1992, Ice-wedge casts and pingo scars: evidence for former permafrost in Connecticut: Wesleyan University, Middletown, CT, unpublished senior thesis, 164 p.

O'Leary, D. W., 1975, Surficial Geologic Map of the Moodus Quadrangle, Connecticut, U.S. Geological Survey Geologic Quadrangle Map GQ-1205, scale 1:24,000.

Oldale, R. N., 1974, Geologic Map of the Dennis Quadrangle, Barnstable, County, Massachusetts: U.S. Geological Survey Geologic Quadrangle Map GQ-1114, scale 1:24,000.

Ridge, J. C. and Larsen, F. D., 1990, Re-evaluation of Antev's New England varve chronology and new radiocarbon dates of sediments from glacial Lake Hitchcock: Geological Society of America Bulletin, v. 42, p. 889-899.

Schafer, J. P., 1968, Periglacial features and pre-Wisconsin weathered rock in the Oxford-Waterbury-Thomaston area, western Connecticut: in Orville, P.M. (ed.), Guidebook for fieldtrips in Connecticut, New England Intercollegiate Geological Conference 60th Annual Meeting, New Haven, CT; Connecticut Geological and Natural History Survey Guidebook no. 2 , p. 1-5.

Schafer, J. P. and Hartshorn, J. H., 1965, The Quaternary of New England: in H.E. Wright, Jr. and Frey, D.G. (eds.). The Quaternary of the United States: Princeton, N.J., Princeton University Press, p. 113-128.

Schile, C. A., 1991, Stationary dune growth in the Connecticut River valley (Springfield South Quadrangle); University of Connecticut, Storrs, CT, unpublished master thesis, 82 p.

Sirkin, Les, 1982, Wisconsinan glaciation of Long Island, New York, to Block Island: in Larsen, G.J. and Stone, B.D., eds., Late Wisconsinan glaciation of New England: Dubuque, Iowa, Kendall/Hunt Publishing Company, p. 35-59.

Stone, B. D., 1978, Surficial Geologic map of the Jewett City Quadrangle, New London County, Connecticut, U.S. Geological Survey Geologic Quadrangle Map GQ-1434, scale !:24,000.

Stone, B. D. and Borns, H.W., 1986, Pleistocene glacial and interglacial stratigraphy of New England, Long Island, and adjacent Georges Bank and Gulf of Maine: in Sibrava, Vladimir, Bowen D.Q., and Richmond, G.M., eds., Quaternary Glaciations in the Northern Hemisphere: Quaternary Science Reviews, v. 5, p. 39-52.

Stone, B. D., Lapham, W. W, and Larsen, F. D., 1992, Glaciation of the Worcester Plateau, Ware-Barre area, and evidence for the succeeding late Woodfordian periglacial climate; Trip C-7, this volume.

Stone, J. R. and Ashley, G. M., 1989, Fossil Pingos?: New evidence for active permafrost in southern New England during Late Wisconsinan deglaciation: Geological Society of America Abstracts with Programs, v. 21 (2), p. 69.

Stone, J. R., Ashley, G. M. and Peteet, D. M., 1991, Cross section through a post-lake Hitchcock surface depression: New ^{14}C dates and evidence for periglacial origin: Geological Society of America, Abstracts with Programs, v. 23 (1), p. 135.

Stone, J. R., Schafer, J. P., and London, E. H., 1982, The surficial geologic maps of Connecticut illustrated by a field trip in central Connecticut; in Guidebook for fieldtrips in Connecticut and south-central Massachusetts, Guidebook No. 5, Connecticut Geological and Natural History Survey, Hartford, Connecticut, p. 5-30.

Stone, J. R., Schafer, J. P., London, E. H., and Thompson, W. B., 1992, Surficial Materials Map of Connecticut: U.S. Geological Survey Special Map; scale 1:125,000.

Stone, J. R. and Schafer, J. P., (in preparation) Quaternary Geologic Map of Connecticut, U.S. Geological Survey Special Map; scale 1:125,000.

Svensson, H., 1969, A type of circular lakes in northernmost Norway: Geografiska Annaler 51A, p. 1-12.

Thorson, R. M. and Webb, R. S., 1991, Postglacial history of a cedar swamp in southeastern Connecticut: Journal of Paleolimnology v. 6, p. 17-35.

Thorson, R. M., Clayton, W. S. and Seeber, Leonardo, 1986, Geologic evidence for a large prehistoric earthquake in eastern Connecticut: Geology, v. 14 (6), p. 463-467.

Vandenberghe, J., 1988, Cryoturbations in Clark, M.J. (ed.), Advances in periglacial geomorphology, John Wiley & Sons Ltd. p. 179-198.

Van Vliet-Lanoe, Brigitte, 1988, The significance of cryoturbation phenomena in environmental reconstruction; Journal of Quaternary Science, v. 3, p. 85-96.