Sub-till Lacustrine Deposits of the Connecticut Valley of Southern New Hampshire and Vermont [Draft Paper]

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Introduction

The Connecticut Valley of southern New Hampshire and Vermont (Fig. 1) is a north-south trending glacially eroded valley surrounded by a hilly terrain having a relief in total of ~500 m and locally averaging 150-300 m. Tributary stream valleys are formed of segments trending in two primary directions, approximately north-south and also east-west, with abundant evidence of glaciation. The glacial geology of the region is dominated by features formed during the last or late Wisconsinan glaciation (Ridge, 1990, 1999, 2001). Especially prominent are:

1) striations, grooves, and whalebacks from a regional ice flow direction of S45°E to due S that overtopped hills in the area and younger flow directions ranging to S40°W from locally-directed lobate ice flow in valleys during deglaciation,
2) gray, stony to sparsely stony till, sometimes in the form of drumlins and thick valley-side till benches that are the eroded remnants of till that once filled some valleys (Ridge, 1988),
3) stratified sand and gravel deposits of tributary proglacial lakes and streams formed during deglaciation,
4) sand and gravel deposits of ice-marginal and meteoric deltas deposited in glacial Lake Hitchcock in the Connecticut Valley during ice recession, and
5) varved fine sand, silt, and clay deposits of the floor of Lake Hitchcock.

Like the rest of New England this area was glaciated during pre-Wisconsinan time (Schafer and Hartshorn, 1965; Stone and Borns, 1986; Oldale and Colman, 1992) but it has not been possible in the study area to confidently identify glacial deposits of pre-Wisconsinan age based on criteria such as weathering characteristics, stratigraphic position, and structural relationships (Oldale and Eskenasy, 1983; Koteff and Pessl, 1985; Newman and others, 1990), or numerical dating techniques (Oldale, 1982; Oldale and others, 1982). In some relatively steep-sided east-west trending segments of tributary valleys surficial units beneath late Wisconsinan till may be preserved where glacial scour did not remove them from their somewhat protected position in valley bottoms. Exposures of these units have not been found in north-south tributary segments which were more deeply scoured during glaciation.

Units beneath Wisconsinan till include highly weathered bedrock (Ridge, 1990, 1999, 2001; i.e. rotten stone, grus, and saprolite) as has been found in other areas of New England and Quebec (Goldthwait and Kruger, 1938; Schafer and Hartshorn, 1965; LaSalle and others, 1985; J.P. Schafer, pers. com.). In the Cold River and Warren Brook valleys stream terrace gravel and colluvium composed of highly weathered rock debris are also found beneath non-weathered late Wisconsinan till (Ridge, 1988). In some cases the colluvium may have a small reworked, pre-weathered glacial component and the stream gravels, judging from their erratic content, seem to have at least part of their source in eroded glacial sediment. In addition, packages of lacustrine sediment composed of laminated fine sand and varved and non-varved, micaceous silt and clay,

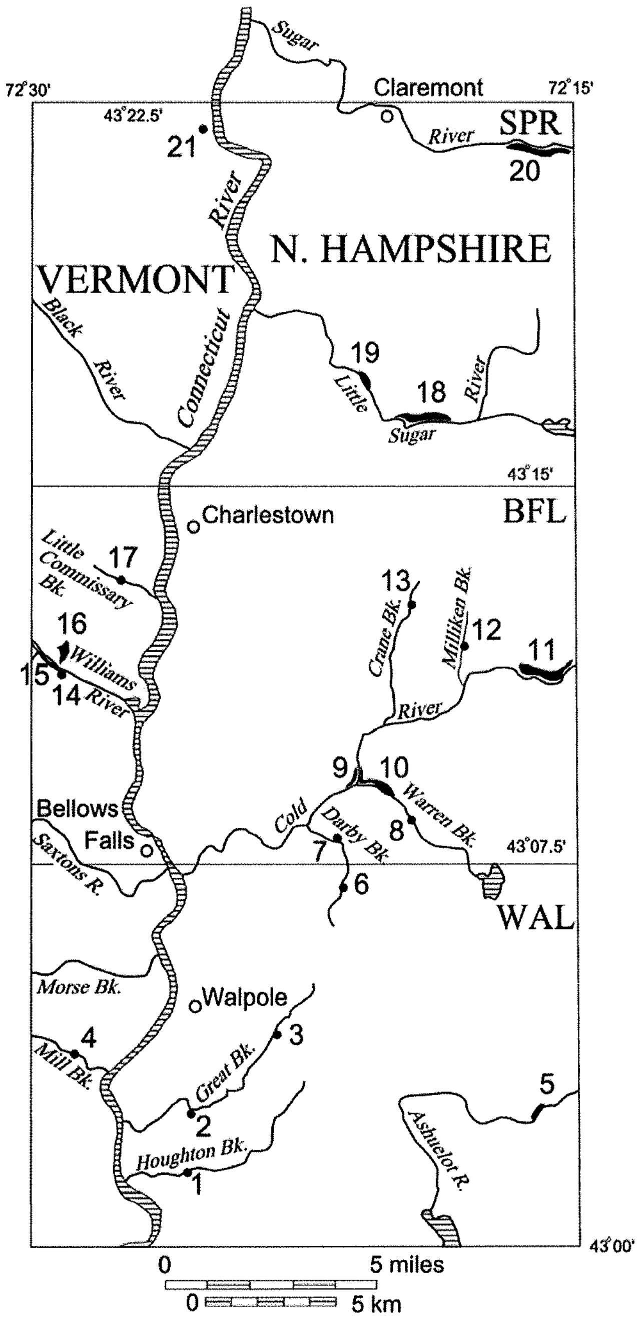


Figure 1. Locations of sub-till lacustrine sediment sites in the Connecticut Valley of southern New Hampshire and Vermont in the Walpole (WAL), Bellows Falls (BFL), and Springfield (SPR) USGS 7.5 x 15-minute quadrangles. See Table 1 for a listing of the sites.

sometimes interbedded with till-like diamicton beds, are found beneath non-weathered till from the last glaciation (Fig. 1, Table 1). Most of the lacustrine deposits have high silt contents, probably derived from glacial rock flour, and an absence of organics. Therefore, with the possible exception of as yet undiagnosed landslide blockage, all of the sub-till lacustrine deposits appear to represent lakes impounded by glacial ice. The sub-till lacustrine units are the subject of this paper and provide us with information about subglacial conditions during the last glaciation as well as the overall characteristic of landscape development during glaciation in this part of New England.

What are Sub-till Lacustrine Deposits?

Sub-till lacustrine deposits are lake deposits found beneath subglacial till deposited during the last glaciation. All of the sub-till lacustrine units appear to be purely clastic with no organic components discernable in outcrop exposures. This supports a glacial origin but it has also been a source of disappointment since the radiocarbon ages of fossils might shed some light on the minimum age of the lacustrine sediment and the age of past glaciations. None of the sub-till lacustrine units are weathered, except by weathering during the current interglacial. However, in some places the oxidation of iron from seeping groundwater with relatively high dissolved ferrous iron concentrations can produce iron staining or cements that may at first glance mimic weathering. The basal units of some of the lacustrine beds, for example along Warren Brook (Fig. 2 and Table 1), may also have units composed of brown to orange sediment reworked from older weathered deposits that are interbedded with gray non-weathered beds. All sub-till lacustrine units are extremely compact as a result of compression beneath overriding ice or burial beneath a thick overburden composed of till that is sometimes overlain by stratified glacial deposits. The tops of sub-till

Table1. Sub-till lacustrine sediment exposures. Highlighted sites have accompanying paleomagnetic analysis.

Site no.	Place	Thickness (m)	Lithologies[1]	Lake type	Age[2]
1	**Houghton Brook**	2.5	md	subglacial cavity	late Wisconsinan
2	lower Great Brook	1.0	md	subglacial cavity	late Wisconsinan
3	upper Great Brook	5.0	mdg	subglacial cavity	late Wisconsinan
4	Mill Brook	1.3	md	subglacial cavity	late Wisconsinan
5	Ashuelot River	2.0	vm	proglacial	late Wisc. advance
6	**upper Darby Brook**	2.0	vmd	proglacial	late Wisc. recession
7	lower Darby Brook	5.0	mdgx	subglacial cavity	late Wisconsinan
8	**Warren Brook**	1.7	vmd	proglacial	late Wisc. advance
9	Cold River	4.0	s	proglacial	late Wisc. advance
10	**Warren Brook - Kmiec**	8.0	vms	proglacial	late Wisc. advance
11	South Acworth	30.0	s	proglacial	late Wisc. advance
12	Milliken Brook	1.7	sd	subglacial cavity	late Wisconsinan
13	Crane Brook	2.0	smd	subglacial cavity	late Wisconsinan
14	Williams River	2.7	sm	proglacial	late Wisc. advance
15	Williams River	10.0	vmd	proglacial	late Wisc. advance
16	Williams River	1.0	md	proglacial	late Wisc. advance
17	Little Commissary Bk.	1.0	md	proglacial	late Wisc. advance
18	Little Sugar River	1.3	vm	proglacial	late Wisc. advance?
19	**Little Sugar River**	1.7	vm	proglacial	late Wisc. advance?
20	**Sugar River**	12.0	vmd	proglacial	late Wisc. advance
21	Blood Brook	4.0	md	subglacial cavity	late Wisconsinan

[1]Sediment types: m = laminated fine sand, silt and clay; s = sand; g = rounded gravel; d = diamicton beds; x = angular rock debris; v = varved sediment.

[2]Approximate age assignments (cal yr BP): late Wisc. advance = ~25,000-40,000 yr BP; late Wisconsinan = period of Late Wisconsinan ice cover = prior to ~25,000 yr BP to no later than ice recession at 14,500 yr BP; late Wisc. recession = 15,100-14,500 yr BP.

lacustrine units are generally deformed to some extent by the shear imparted by overriding ice. In addition, overlying till is usually partly composed of deformed lacustrine sediment. The lower parts of the lacustrine units sometimes escaped deformation except for compaction. In many places sub-till lacustrine sediment extends beneath the floors of modern valleys indicating that the ancient lakes they represent were in valleys that were deeper than modern valleys.

The sub-till lacustrine units appear to represent deposition in relatively small tributary lakes. All sub-till lacustrine units occur in tributary valleys (Fig. 1 and Table 1) usually 10's of meters above the surface elevation of Lake Hitchcock that formed during the last deglaciation in the Connecticut Valley. It also appears that the advance of the last glacier in the Connecticut Valley occurred at a time when there was a river occupying the valley and not a lake. Evidence of this river occurs in outcrops with till overlying fluvial gravel in West Lebanon, NH (see Stop 1 of Larson, 1987) and in subsurface borings in Putney, VT where gravel is overlain by more than 30 m of till (Ridge, 1990). This seems to eliminate them as being deposits from the tributary extensions of large proglacial lakes in the Connecticut Valley proper either during ice recession or advance of the last or earlier glaciations.

There are two major types of sub-till lacustrine deposits in the Connecticut Valley field area, the deposits of ice-marginal proglacial lakes and subglacial cavities, both primarily

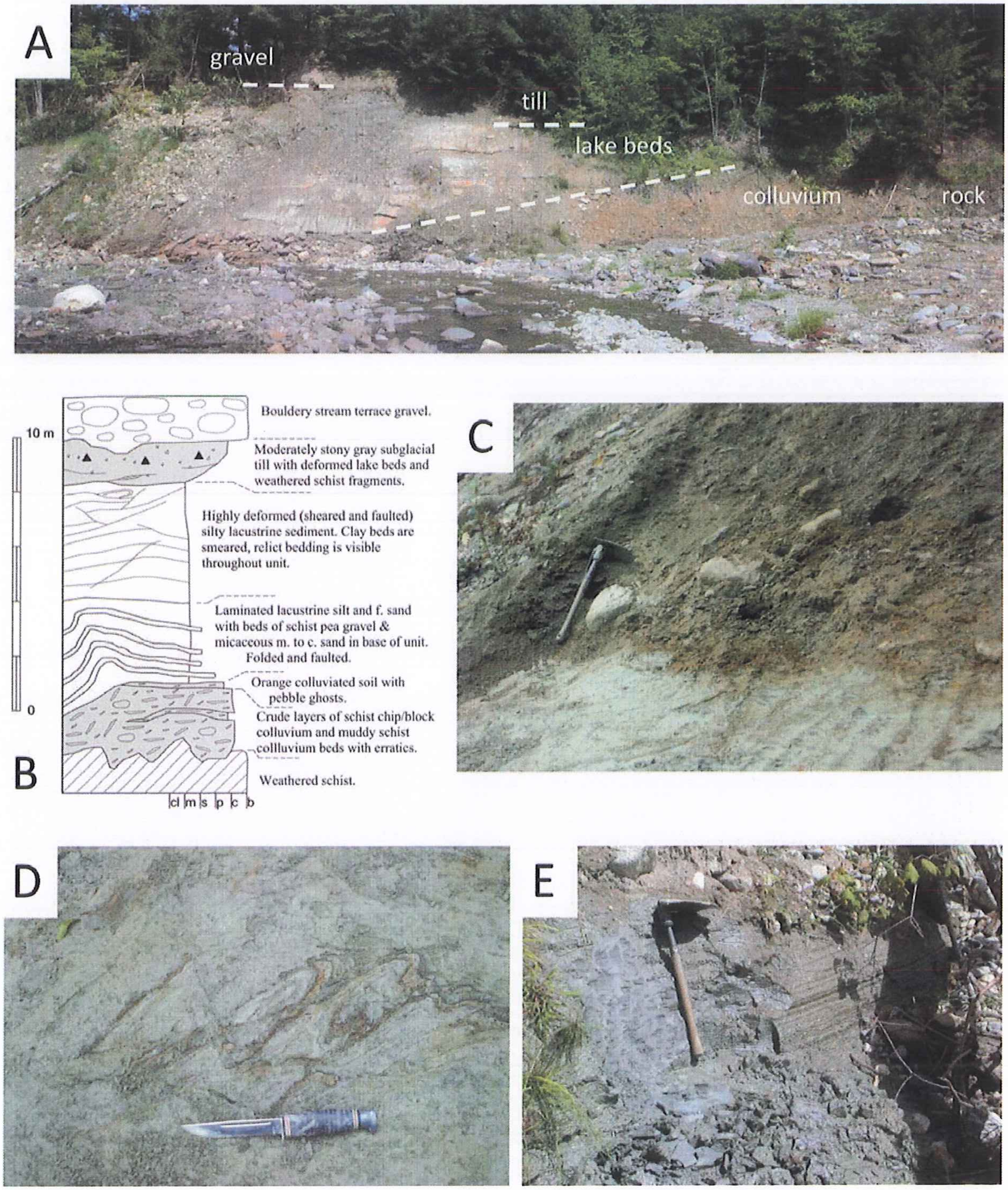


Figure 2. The Warren Brook (Kmiec) section (site 10) near the confluence of the Cold River and Warren Brook in Alstead, NH. A. Panoramic view of section. B. Stratigraphic column and description of section. C. Close-up view of contact of late Wisconsinan till resting on sub-till lacustrine sediment. Shovel is 55 cm long. D. Subglacial deformation (faults and associated tight drag folds) in the upper deformed part of the sub-till lacustrine sediment. Knife is 22 cm long. E. Undeformed sandy and silty varves in the sub-till lake beds 100 m upstream from the main section shown in the panoramic view. Shovel is 55 cm long.

associated with late Wisconsinan glaciation. The two types of deposits and their ages are distinguished from each other based on combined lithologic, stratigraphic, and paleomagnetic characteristics.

Paleomagnetic Methodology and General Results

Paleomagnetic directions in fine-grained clastic sediment represent a time-dependent signal that can be used to test hypotheses regarding the ages of deposits by either: 1) disproving an hypothesized age for a site when it has magnetic directions that are inconsistent with known paleomagnetic directions for a particular time, or 2) disproving that two sites have the same age when they have paleomagnetic directions that are different (Ridge and others, 1990). Unfortunately paleomagnetic measurements cannot be used to prove the ages of deposits or their contemporaneity since paleomagnetic directions have a secular variation and may be repeated through time. The hypothesis that sub-till lacustrine beds may be associated with late Wisconsinan ice recession is an hypothesis easily tested since the paleomagnetic declination record of the time of late Wisconsinan ice recession in southern New Hampshire and Vermont (15,100-14,500 cal yr BP) is well known from the analysis of varves of glacial lakes Hitchcock and Merrimack (Johnson and others, 1948; Verosub, 1979; Ridge and others, 2001; see update by Ridge, 2004, 2010).

Unfortunately not all sub-till lacustrine deposits and paleomagnetic parameters are acceptable for stratigraphic analysis using paleomagnetism. Deposits must be fine-grained and have sufficient populations of magnetic particles in the clay to very fine silt size that are free to magnetically orient to the direction of the ambient geomagnetic field during deposition. Luckily, laminated muddy glacial lacustrine sediments in the Connecticut Valley have abundant magnetite with the requisite grain size. An additional concern with sub-till lacustrine units is that they can be deformed by either pervasive shearing or rotation by overriding ice, thus disturbing magnetic directions acquired during deposition. At some sites bedding may be intact but vertical and horizontal compression may combine to create a spaced cleavage, which can be associated with disturbed magnetic signals (Ridge and others, 1990). At most of the 21 sites studied for this paper the upper parts of lacustrine units were too deformed for meaningful paleomagnetic analysis but at six of the sites the lower parts of laminated fine-grained sediment escaped deformation (Table 1).

The paleomagnetic parameter most useful for chronostratigraphic analysis of glacial lacustrine deposits is remanent declination because it is locked into the sediment within a few years of deposition and is not significantly disturbed by post-depositional compaction. Therefore, it is a valuable recorder of the declination of the geomagnetic field at the time of deposition and the secular variation of the geomagnetic field (Johnson and others, 1948; Graham, 1949; Granar, 1958; Noel, 1975; Verosub, 1975). Remanent inclination varies over a narrower range of values than declination and can be inconsistent in deposits of similar age because of depositional processes (Johnson and others, 1948; King, 1955; Granar, 1958; Griffiths and others, 1960; Hamilton and King, 1964; Verosub, 1977; Ridge and others, 1990) and flattening during compaction (Blow and Hamilton, 1978; Anson and Kodama, 1987). The magnetic intensity of glacial lake sediment can also vary with secular variations in the intensity of the geomagnetic field. However, the magnetic moments and intensities measured in glacial lake deposits of the Connecticut Valley vary over 2 orders of magnitude in response to differences in grain size and magnetite concentrations. These variations are hard to separate from smaller variations associated with geomagnetic field intensity in a way that produces a level of precision useful for stratigraphic analysis. None of the late Pleistocene sediments studied in the Connecticut Valley are old enough to have been influenced by reversals in geomagnetic polarity.

Ten to four oriented blocks were collected from each of 2-5 horizons at outcrops of undeformed sediment for paleomagnetic analysis. These blocks were then trimmed to fit in 2.5-cm long x 2.54-cm OD plastic tubes and sealed with clear 5-minute epoxy (see Ridge, 2010 for method). Remanent magnetization (remanence) of the samples was measured using a Molspin Minispin spinner magnetometer. After measuring the natural remanence of all samples pilot samples were sequentially subjected to alternating field (AF) demagnetization using a Sapphire Instruments SI-4 single axis demagnetizer. Steps of 2.5, 5, 10, 20, 30, 40, 50, 60, 70, 80, 100, and 120 mT were used to determine the optimum AF demagnetization treatment for removing viscous remanent magnetization (VRM) from the rest of the samples. Directions of the pilot samples remained unchanged after AF demagnetization at 20-30 mT and the samples had median destructive fields of 25-60 mT. All remaining samples were AF demagnetized at 30 mT for removal of VRM (Table 2 and Fig. 3).

Table 2. Paleomagnetic data from sub-till lacustrine sediment sites. See Figure 1 and Table 1 for locations.

Site no. - place	Samples (n)/ horizons	30 mT dec.(°)	30 mT inc.(°)	K / α_{95}(°)[1]	AMS principal axes:	k_{max}	k_{int}	k_{min}
					k	5.07	5.01	4.33
1 - Houghton Bk.	20/5	358.1	13.1	419 / 1.6	dec.(°)	336	66.1	207.4
					inc.(°)	2.8	2.8	86.1
					k	4.08	4.04	3.63
6 - upper Darby Bk.	16/2	338.0	42.4	507 / 1.6	dec.(°)	260.2	350.3	168.6
					inc.(°)	0.1	5.5	84.7
					k	4.01	3.95	3.72
8 - Warren Bk.	18/2	329.0	51.8	411 / 1.7	dec.(°)	229.2	319.3	112.5
					inc.(°)	2.7	3.3	87
10 - Warren Bk.					k	11.39	11.16	9.51
(Kmiec section)	16/4	347.1	34.7	615 / 1.5	dec.(°)	16.1	285.9	117.2
					inc.(°)	1.2	6.7	83.2
					k	8.26	8.13	7.34
19 - Little Sugar R.	25/4	8.2	26.8	275 / 1.8	dec.(°)	333.2	63.7	190.5
					inc.(°)	5.5	3.9	83.6
					k	5.16	5.05	4.28
20 - Sugar River	16/4	23.0	21.4	244 / 2.4	dec.(°)	29.2	299.2	206.7
					inc.(°)	4.9	0.7	84.8

[1]Statistical parameters for remanent magnetization: K = cluster coefficient; α_{95} = 95% cone of confidence.

Anisotropy of magnetic susceptibility (AMS) was measured to determine whether coarse, multi-domain particles in the samples had been preferentially aligned by currents or mechanical deformation processes to a direction that might coincide with remanence directions. Coincident remanence and maximum susceptibility directions may indicate that remanence directions were fixed by forces other than the geomagnetic field at or after the time of deposition. Bulk (Z-axis) susceptibility magnitudes were determined using a Molspin bulk susceptibility device and principal AMS axes were determined using a Molspin Minisep AMS device. All samples displayed nearly horizontal maximum and intermediate susceptibility axes with nearly similar magnitudes while the near vertical minimum axes had magnitudes approximately 10-20% lower

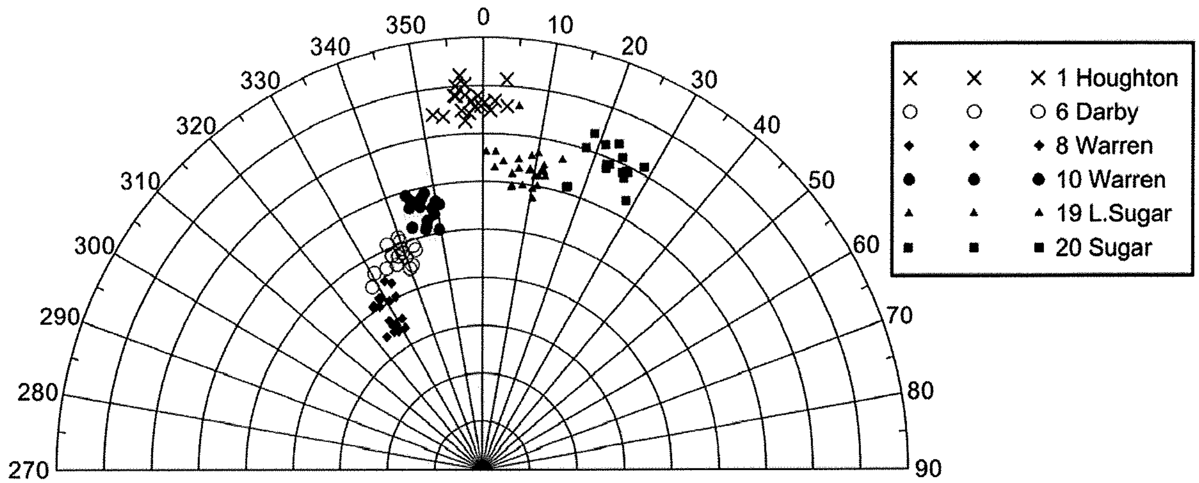
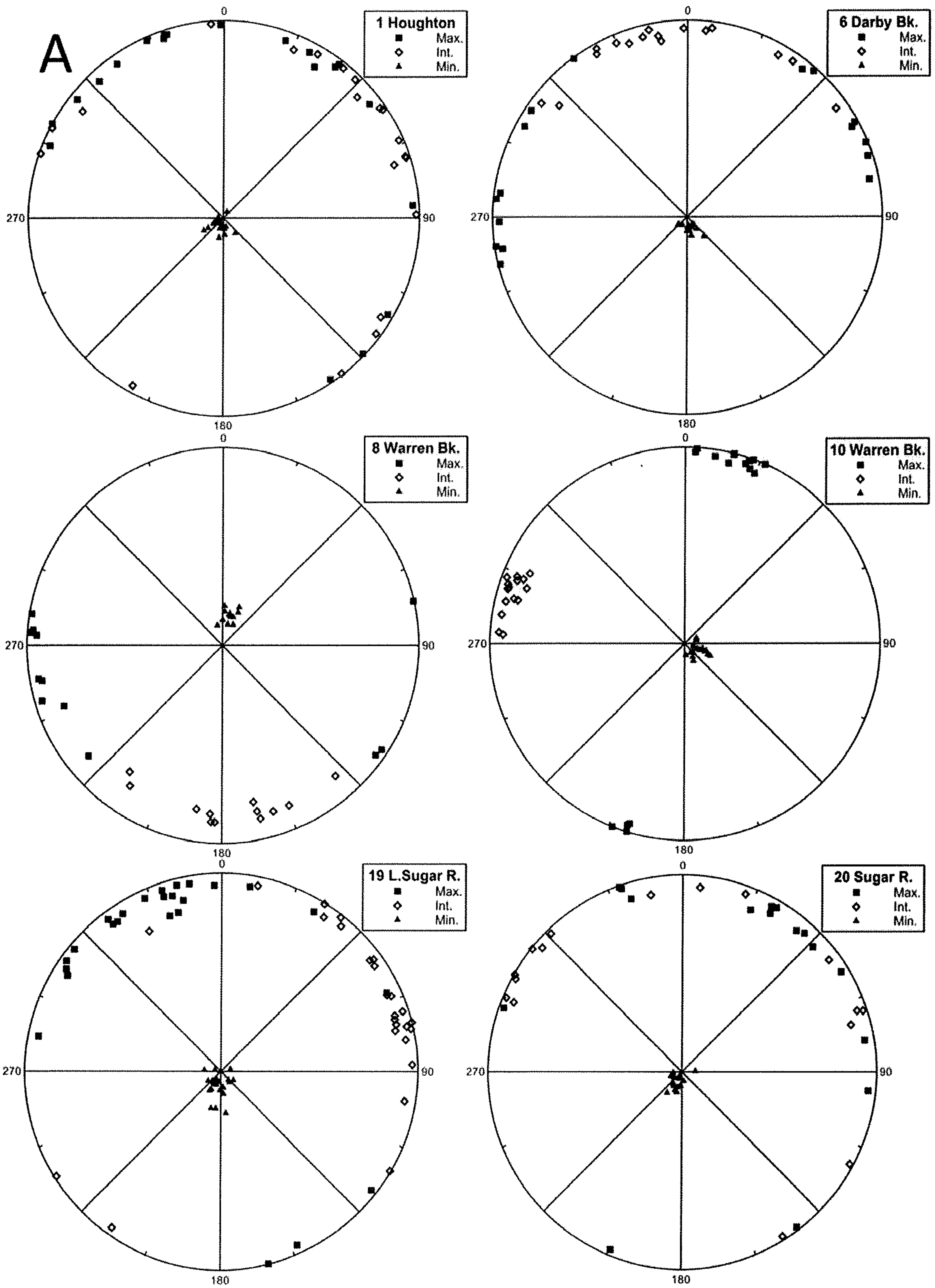


Figure 3. Polar plot of remanent magnetization directions in sub-till lacustrine deposits after AF demagnetization at 30 mT.

resulting in a well defined horizontal foliation (Fig. 4 and Table 2). Maximum susceptibility directions were also generally scattered in the horizontal and showed no correspondence to remanence directions. The one exception to this general situation was at one of the Warren brook sites (site 10, Fig. 1, Tables 1 and 2). At this site maximum susceptibility axes clustered but in the horizontal with an azimuth (0-25°) far removed from the remanent declination (347.1° $\pm$ 1.5°). In addition the maximum and intermediate susceptibility axes were very close to each other in magnitude indicating a very weak lineation. The AMS configurations are consistent with sediment that was compacted but not horizontally deformed and in which remanence directions record the geomagnetic field at the time of deposition.

Sub-till Lacustrine Deposits of Ice-marginal Proglacial Lakes

The most widespread occurrences of sub-till lacustrine sediment appear to have been deposited in proglacial lakes impounded in east-west tributary segments during the advance of the last glacier. Key examples are in the Williams, Sugar, and Cold River and Warren Brook valleys (Figs. 1, 2, 5, 6, and 7; Table 1). These lakes, here called "advance lakes", were formed when the advancing lobate glacier front blocked down valley sections of tributaries closer to the Connecticut Valley and impounded lakes in up valley areas that would later be overrun by ice. The deposits of these lakes are buried beneath till and tend to be varved where finer-grained facies occur, but there may also be thick and extensive areas of well-sorted fine sand representing prodeltaic areas at the mouths of outwash streams. In the Cold River Valley (sites 9 and 11 on Fig. 1 and Table 1) sand was deposited from up valley sources and grades into muddy varves in the Warren Brook tributary valley (sites 8 and 10, Figs. 1, 2, and 7, Table 1) which was a quieter environment probably not yet receiving glacial meltwater in its headwaters. Varved units have well defined winter (non-melt season) layers that are more than 90% clay and sandy to silty summer (melt season) layers composed of multiple micro-graded units (Fig. 2E, 7A, and 7B). All muddy units in sub-till deposits representing proglacial lakes preserve trace fossils of the ichnogenus *Cochlichnus* (Fig. 8), probably formed by the crawling of either nematode worms or insect larvae. In the case of the Sugar River Valley (site 20, Fig. 1 and Table 1), varved lake beds show evidence of ice rafting in the form of diamicton pellets and dropstones and also


A
0
90
180
270
1 Houghton
Max.
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6 Darby Bk.
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8 Warren Bk.
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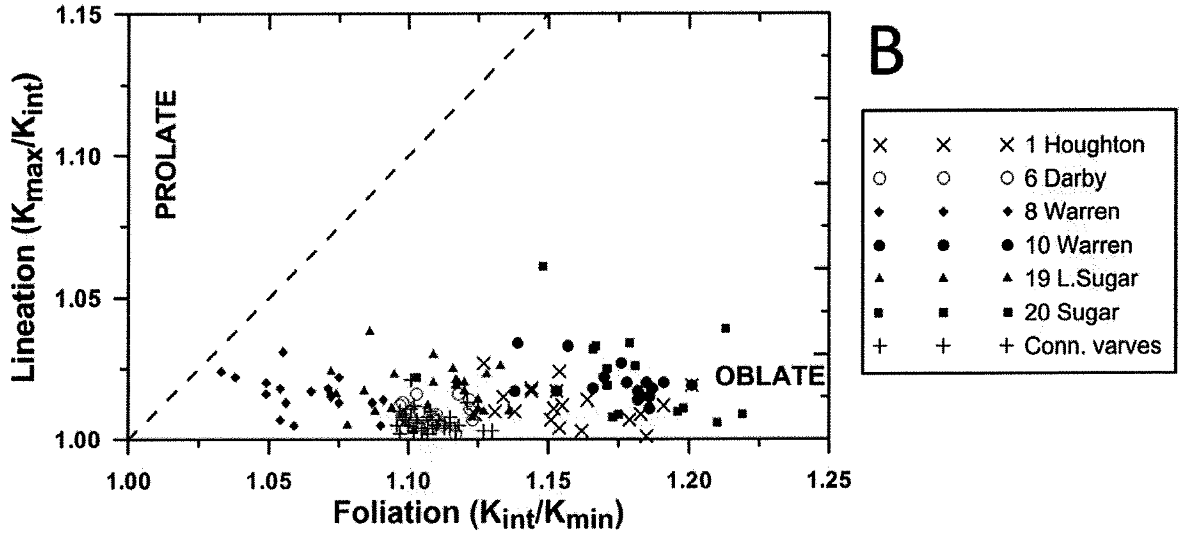

Figure 4. AMS data for sub-till lacustrine deposits. A. Polar plots of principal axes of AMS. B. Flinn plot (ratios of k_{max}/k_{int} vs. k_{int}/k_{min}) showing foliation and lineation characteristics of AMS. Varves from South Windsor, Connecticut are shown for comparison to varves not compressed by overriding ice.

Figure 5. Stratigraphic section in the Williams River Valley (site 14 on Fig. 1 and Table 1) showing late Wisconsinan till overlying deformed, varved sub-till lacustrine sediment.

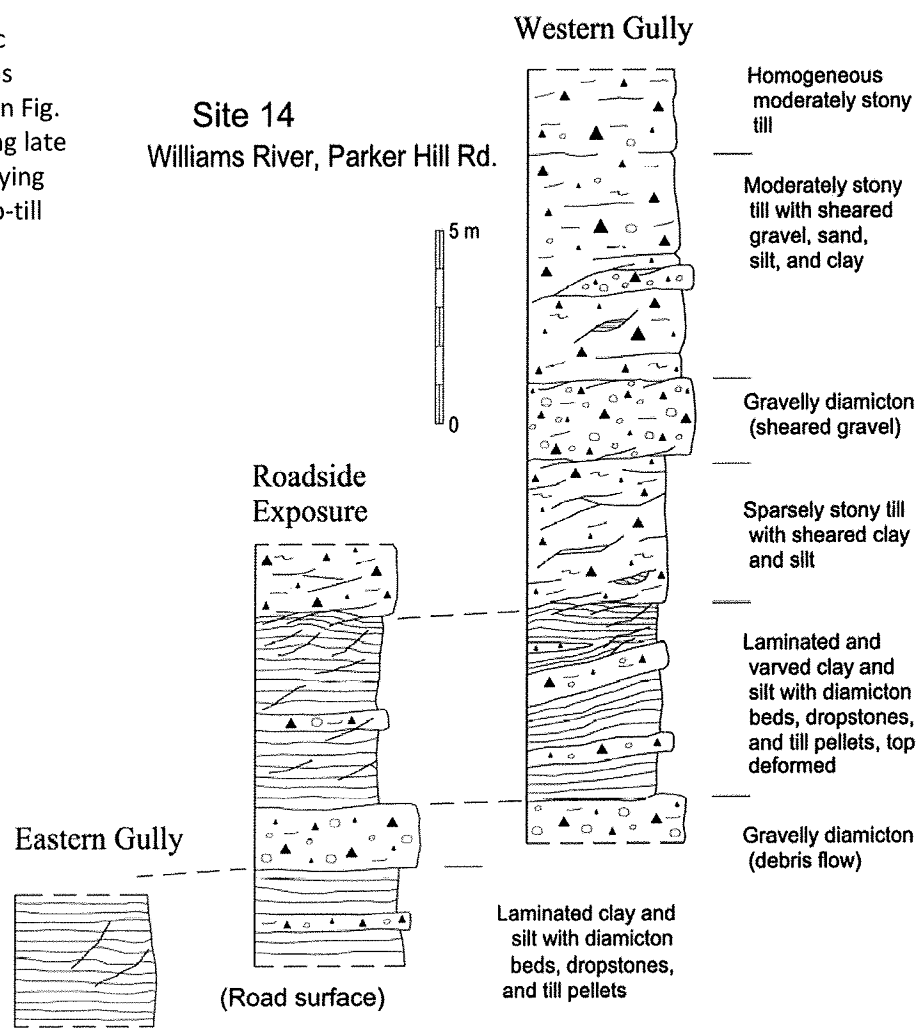

Figure 6. Varved sub-till lacustrine beds along the south side of the Sugar River near Claremont, NH. These units have high concentrations of diamicton pellets and dropstones and are interbedded with many diamicton beds representing glacial debris flows. Shovel is 55 cm long.

interbed with diamicton beds that appear to represent debris flows. Both the diamicton pellets and beds have exactly the same lithology as the overlying till. Deposition in this valley may have been in a deep lake that promoted calving and an environment that was ice-proximal. The local topography lends itself to the simultaneous advance of ice into the valley from both its eastern and western ends and it is possible there was a deep advance lake choked with debris-laden icebergs as the last ice sheet invaded this area.

Ice-proximal varves of advance lakes are thinner and muddier than those deposited in Lake Hitchcock during late Wisconsinan ice recession. Perhaps this is a reflection of smaller meltwater discharges and gentler bottom currents during glacial advance as compared to during ice recession when the ice sheet had a negative mass balance and a much higher meltwater discharge. The advance lake deposits are also always finer than the deposits of tributary lakes formed during ice recession, which are dominated by deltaic sand and gravel. When recessional deposits have varves they are generally thick sandy couplets. This may be a reflection of higher meltwater discharges during ice recession, but also shallower water columns (above the till units) that would have facilitated higher current velocities and flushing of clay and silt.

From samples of advance lake deposits it has been possible to record changes in paleomagnetic declination that occurred as ice advanced from north to south from the Sugar to Cold River Valley. The oldest advance lake deposits in the Sugar River Valley record a paleomagnetic declination of 20-25° (Table 2, Fig. 3). By the time ice advanced to the Cold River Valley declination had shifted to 345-325°. A declination of 6-10° in the Little Sugar River Valley may represent an intermediate point in this transition but whether this deposit represents an advance lake of the last glaciation remains uncertain. With the addition of more paleomagnetic measurements from advance lake deposits it may be possible to assemble a much longer secular variation record of declination for the period of ice advance of the last glaciation.

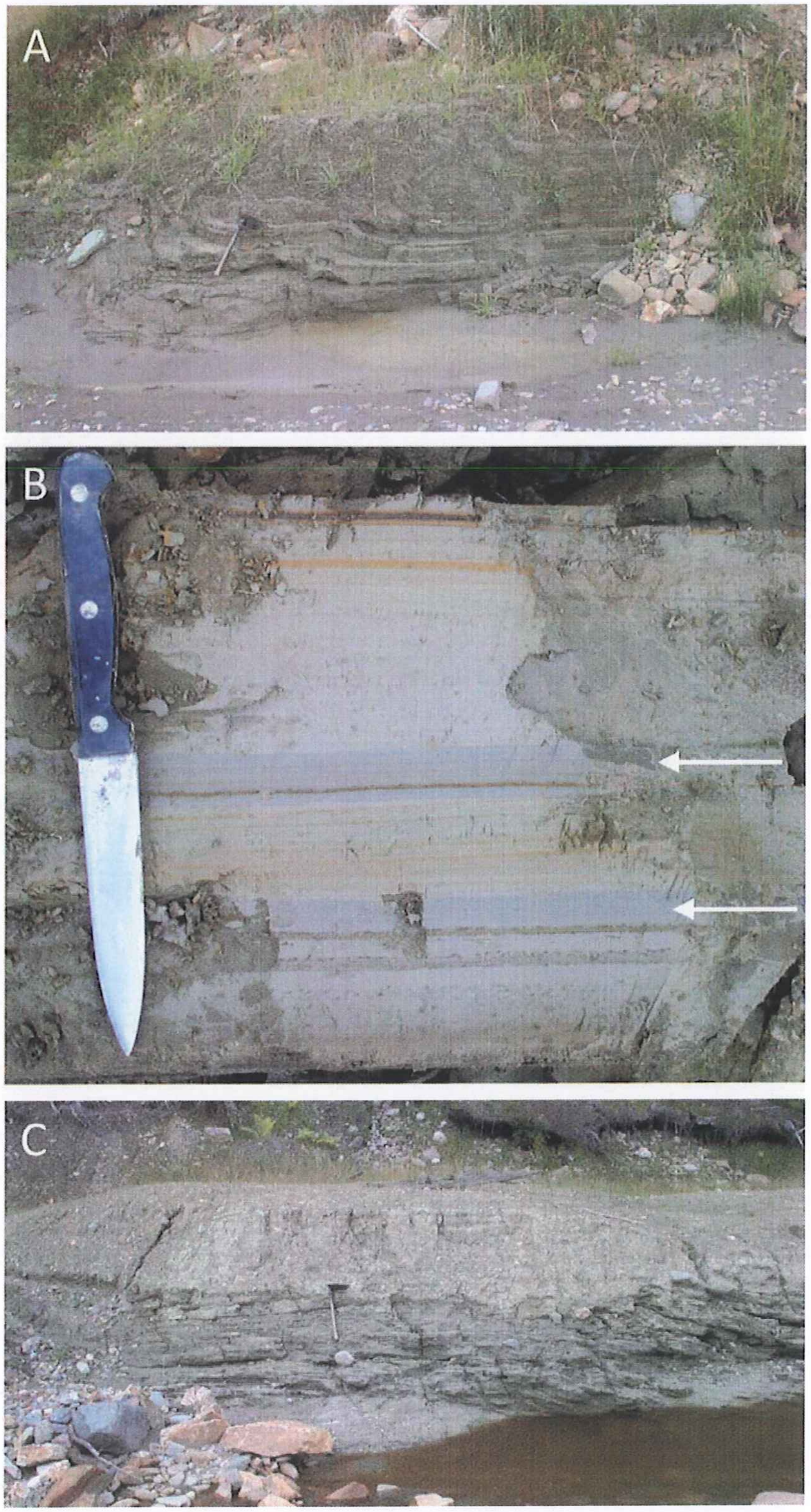

Figure 7. Varved sub-till lacustrine beds from an advance lake in up valley areas of Warren Brook (site 8 on Fig. 1 and Table 1). This area was a tributary extension of a larger lake in the Cold River Valley and was a lower energy environment not yet fed by glacial meltwater in its headwaters. A. Varve section that is very lightly warped and mostly escaped deformation by overriding ice. The section is today truncated by postglacial terrace gravel. Shovel is 55 cm long. B. Varves in the Warren Brook section showing thick discrete winter clay beds (arrows). Knife is 25 cm. C. Upper deformed part of Warren Brook section with laminated silt and clay interbedded with diamicton layers. Shovel is 55 cm long.

The possibility that some of the proglacial lake deposits represent lake beds buried by till of minor readvances during ice recession should also be considered. This scenario is not possible for some of the advance deposits, for example in the Warren Brook and Cold River valleys

Figure 8. Sinusoidal crawling traces of the ichnogenus *Cochlichnus* commonly found in muddy units of sub-till lacustrine beds formed in proglacial lakes. Example shown here is from late Wisconsinan varves deposited in Lake Hitchcock at South Windsor Connecticut. Scale in cm.

(Ridge, 1988, 1999), where the bottom of the lacustrine unit drapes preglacial substrates such as highly weathered rock or colluvium containing weathered debris (Figs. 1 and 2). Substrates that are glacial surfaces, such as till, non-weathered striated rock, or ice-proximal stratified deposits, would be a requirement beneath lake beds formed during ice recession. Unfortunately the bottoms of many sub-till lacustrine units are not exposed or occur beneath modern valley bottoms and the character of this lower contact often remains hidden. Their burial beneath very thick till (>10 m) and extreme compaction suggests that they may have formed prior to a full glaciation and not simply buried by till of a minor readvance.

Paleomagnetic measurements of undeformed sub-till units provide an additional test of whether lake beds were buried by till of a late Wisconsinan readvance. So far only sub-till lake beds in the upper Darby Brook Valley (Figs. 1 and 9, Table 1) overlie fresh bluish gray till and have a paleomagnetic declination (338° $\pm$ 1.6°, Fig. 3, Table 2) consistent with the last deglaciation in southern New Hampshire and Vermont, a time (15,100-14,500 cal yr BP) for which the paleomagnetic declination record is well known (320-340°) from varves of Lake Hitchcock (Ridge, 2003, 2004). The lacustrine sediment here is varved (less than 5 years), interbedded with sparsely stony diamicton beds, displays the ichnogenus *Cochlichnus*, and is overlain by sparsely stony and silty till containing reworked lacustrine sediment.

An additional consideration is whether some of the lake beds attributed to advance lakes could represent lakes formed during the recession of pre-Wisconsinan ice. Three things seem to be inconsistent with this hypothesis. First, like lake beds put down just prior to a minor readvance during the last deglaciation they would have to drape a glacial substrate. This would automatically eliminate the lake deposits in the Cold River and Warren Brook valleys (sites 8 and 10, Figs. 1, 2, and 7, Table 1) where they bury a non-glacial substrate composed of highly weathered rock and colluvium containing weathered debris. Second, recessional lake deposits from pre-Wisconsinan deglaciations like those of the last deglaciation, should be dominated by sand and gravel from lakes that were relatively shallow and small and had strong currents flushing clay and silt. Finally, while the absence of weathering is not conclusive, the complete absence of ancient weathering in any of the sub-till lacustrine units suggests formation after all major pre-Wisconsinan interglacial periods. There is one possible exception, however, in the

Figure 9. Sub-till lacustrine beds along upper Darby Brook that are sandwiched between stony till below and sparsely stony and silty till above. The section appears to represent a minor late Wisconsinan readvance. Shovel is 55 cm long.

Little Sugar River Valley. The sub-till lacustrine unit at this site (site 19, Fig. 1 and Table 1) is varved fine sand, silt, and clay, deformed in the upper part of the unit and overlain by sparsely stony, clayey and silty till. The lower contact of this unit is concealed by thick slump debris but bluish-gray unoxidized till is exposed in the stream a few meters below. The outcrop is suggestive of a late Wisconsinan readvance, a hypothesis that is consistent with local evidence of small readvances that constructed end moraines in the area between Charlestown and Claremont, NH (Ridge 2001, 2004). However, paleomagnetic declinations in the lacustrine unit ($8.2° \pm 1.8°$) are inconsistent with paleomagnetic declinations from the time of late Wisconsinan ice recession (320-340°). Unless a contact of non-glacial origin is exposed between the lake beds and underlying till, as would be expected for an advance lake deposit, or the paleomagnetic declinations are wrong for some unforeseen reason, the sub-till lacustrine unit and the till beneath it may be non-weathered pre-Wisconsinan units.

Sub-till Lacustrine Deposits of Subglacial Cavities

Another major type of sub-till lacustrine unit is deposited in water-filled subglacial cavities. Subglacial cavity deposits, from the size of small seams to large cavities on the lee (south to southeast) sides of bedrock ledges and hills, form where the glacier can bridge a water-filled cavity that is held open by fluid pressure. These deposits are common on a small scale within till exposures in the Connecticut Valley field area. Figure 1 and Table 1 do not reveal the vast number of small (<50 cm thick,<10 m wide) occurrences of this type of deposit in till from the last glaciation, nor do they indicate the many deposits of this type that are potentially concealed along lower valley sides by slump debris. The largest subglacial cavity occurrences are along Houghton Brook, upper Great Brook, lower Darby Brook, and Crane Brook (Fig. 10).

Subglacial cavity deposits all occur within late Wisconsinan till and separate non-weathered till of the same lithology above and below. They can occur at high elevations on the south sides of hills where the occurrence of a proglacial lake from either advance or recession would be hard to explain. In addition they always seem to be extremely compacted, more than the sub-till lacustrine deposits of proglacial origin. These deposits always have large amounts of course debris and till pellets mixed into their fine-grained laminated beds. The debris and pellets are consistent with sediment falling from the ceiling of a subglacial cavity. The cavity deposits are also always interbedded with diamicton beds that appear to represent debris flows. Sorted clayey and silty units may be rhythmic but do not have the nearly pure clay beds that appear in varves of proglacial lakes. The ichnogenus *Cochlichnus* has not yet been found in these deposits and it seems that the occurrence of organisms in subglacial cavities is either extremely rare or not possible. The upper surfaces of cavity lake deposits are only sheared across a few centimeters beneath the till above and only lightly deformed throughout. This could be the result of water pressure in the cavity preventing the full grounding of ice, which then restricts pervasive shearing of the lacustrine sediment.

The Houghton Brook site (1 on Fig. 1, Table 1) has the best example of well laminated deposits of all the subglacial cavity deposits. They are clayey silt and muddy fine sand with abundant diamicton pellets, dropstones, and diamicton beds (Figs. 10A and 10B). The deposits are very similar in structure and lithology to laminated diamictite beds of the Squantum "Tillite" of the Boston Basin that have been attributed to debris flow deposition (Bailey, 1993). The deposit along lower Darby Brook extends for about 100 m and is 5 m of coarse angular schist debris resembling talus at its northern (up ice) end in the lee of a bedrock ledge. It grades southward to bedded diamicton and eventually rhythmic laminated clayey silt and muddy fine sand (Fig. 10D).

The Houghton Brook site (Figs. 1, 10A and 10B, Table 1) was the only cavity lake deposit that could be analyzed for its paleomagnetic characteristics (Table 2) since coarse debris in many of these deposits makes them difficult to sample. The samples at Houghton Brook had very good internal consistency (high cluster coefficient and low α_{95}) like the samples from varved sub-till lacustrine sediment of proglacial origin and recessional varves of many other glacial lakes in New England and New York State (Ridge and others, 1990, 1999, 2001). In addition to having a remanent declination that was distinct from that of the sub-till lacustrine sites remanent inclinations were the lowest of all sample sites ($13.1° \pm 1.6°$) possibly the result of extreme flattening due to compaction by overriding ice.

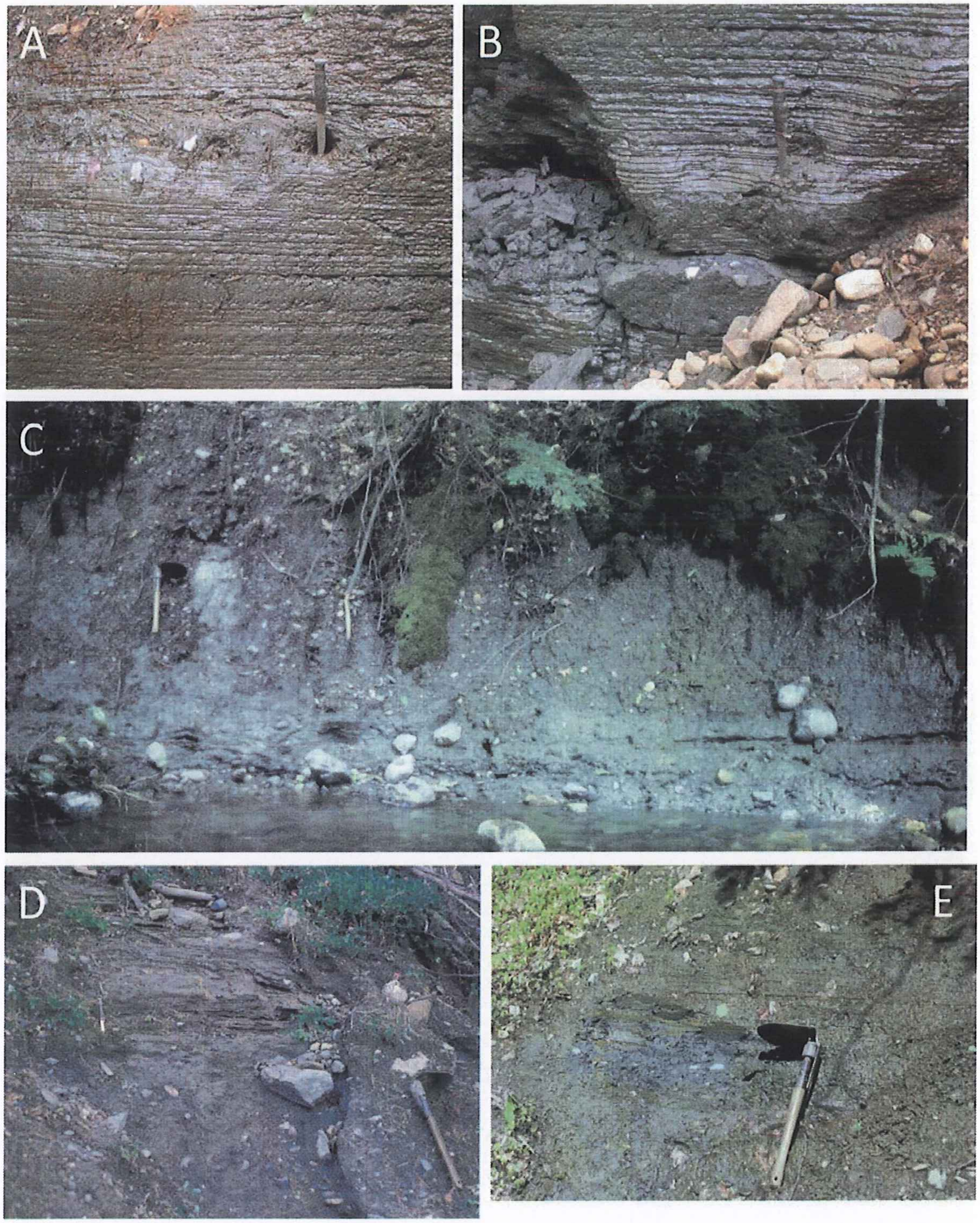

Figure 10. Exposures of sub-till lacustrine deposits from subglacial cavities. A. Interbedded rhythmic, laminated, diamicton pellet-rich mud, coarse sand, and diamicton beds along Houghton Brook (site 1, Fig. 1). B. Thick stony diamicton bed (near base) in laminated stony and pellet-rich rhythmic mud with coarse sand partings along Houghton Brook (site 1, Fig. 1). C. Interbedded laminated mud and coarse diamicton beds along upper Great Brook (site 3, Fig. 1). D. Laminated mud and muddy sand resting on stony till in lower Darby Brook section (site 7, Fig. 1). E. Laminated rhythmic mud, muddy sand, and diamicton beds along Crane Brook (site 13, Fig. 1).

Discussion Point 1: Does Glaciation Cause More or Less Relief?

Several aspects of sub-till lacustrine deposits and their widespread preservation give us clues about the character of ice advance and subglacial bed conditions and erosion during the last glaciation. Sub-till lacustrine deposits from advance lakes, and the preglacial features beneath them, are preserved in east-west trending valley segments. This contradicts the somewhat widely held notion that the glaciation of New England resulted in widespread scour of the landscape except in areas of resistant rock which today stand high creating the relief we see today. To the contrary there are many topographical situations that readily protect and preserve pre-advance sediment and in these areas valleys are not significantly deeper than prior to glaciation. It should be emphasized that this was a conclusion arrived at by Schafer and Hartshorn (1965) nearly 50 years ago. In some situations preglacial sediment and weathering is well preserved in valley bottoms while adjacent hill tops are bare and heavily scoured. This situation in many areas of the Connecticut Valley represents a decrease, not an increase in relief as a result of glaciation.

Discussion Point 2: Erosion Protection and Till Deposition in East-West Valleys

In east-west valley segments there is a higher likelihood of formation of an advance lake because of the more likely down valley impoundment by advancing lobate ice. In north-south tributaries there is nothing to hold the water in the valley unless the valley drains to the north, which is not a common situation in the Connecticut Valley area of southern New Hampshire and Vermont. Ice advancing into a water body, at least initially, would have had a more difficult time eroding the valley bottom substrate due to the partial buoyancy of an advancing ice front. Sub-till lacustrine sediment shielded preglacial substrates (soils and colluvium) at least from some erosion and provided more protection than in a valley that did not have an advance lake. However, this cannot protect pre-advance sediments forever. With continued flow of ice the weak lake sediment substrate may become more eroded as ice thickens. Erosion of the upper surfaces of sub-till units is evident from the deformed tops of most of the advance lake deposits which have been pervasively sheared. However, the sediment is also being compacted as water drains from the sediment under an increasing ice load, allowing shear resistance to build. Sediments dominated by fine sand and silt, common in many advance proglacial lake deposits, will begin to form very resistant substrates because they can drain somewhat easily under overburden pressures, which allows them to compact easily.

Eventually till deposition replaces substrate erosion and sub-till deposits get buried beneath often very thick valley-bottom till. The deposition of thick till in this situation can only be explained if subglacial sediment is mobile, i.e. wet and weak, and readily transported to valley bottoms where it is deposited. The mobility of subglacial sediment is easily seen in the pervasive deformation that occurs in the upper parts of sub-till lacustrine deposits. The lacustrine sediment has its bedded sediment turned into a homogenized sparsely stony deformation till and it gets incorporated into the basal parts of stony till above. Transport of sediment as debris frozen in basal ice of the glacier and later released through basal melting cannot explain the large-scale transport necessary for deposition of thick till. It would require too much flux of heat to melt sufficient basal ice and release the large volume of debris that must accumulate to form thick till.

The mobility of subglacial sediment requires the bed of the glacier to be wet, especially on the north sides of valleys where sediment is transported into valley bottoms. The wet mobile bed being dragged southward into valley bottoms will have a more difficult time leaving the valley by up slope transport. On the south sides of valleys up slope mobility will require higher shear stresses in the sediment than was required to transport it down slope to valley bottoms. A net accumulation of till will occur in valley bottoms if these higher shear stresses are not generated, creating the thick protective till cover we today find over sub-till lacustrine units. Since north-south trending valleys have no defending southern slopes subglacial transport will not be impeded and erosional scour seems like a more viable process when there is not a mechanism to facilitate valley bottom till accumulation.

Discussion Point 3: Water-filled Subglacial Cavities – Are they recessional features?

Subglacial cavity deposits within till of the last glaciation fill small centimeter-thick seams to large cavities several meters high and extending for up to 100 meters. All of the larger cavity deposits discovered so far (Table 1) are associated with deposition on the lee (south to southeast) sides of bedrock ledges and hills. As ice passed over these features it bridged the space on the lee side creating a hollow that became the site of stratified sediment deposition. We do not know the exact environment in which the cavities form relative to the glacier margin but given their wide range of elevations and there occurrence beneath thick till it seems as though relatively thick ice cover, significantly back from the margin of the glacier, is required for their formation. Cavities would have to remain water-filled for prolonged periods (more than a year) to facilitate the uninterrupted deposition of stratified deposits and would have to have water pressures capable of preventing the collapse and grounding of the ice above. Formation within a kilometer of a receding ice margin under relatively thin ice would not provide the time necessary for burial by thick till nor would it be able to maintain water pressures given that areas close to the margin are likely to periodically drain.

Unfortunately there is presently no way to determine when the water-filled subglacial cavities formed. We don't know if they represent the same general time interval or the same general subglacial zone at some consistent distance behind the glacier margin. However, we do know that cavity formation was active at a time of high basal water pressure. Perhaps it represents a time during ice recession when a large volume of meltwater on the surface of the glacier is finding its way to the glacier bed, creating a situation that allows high basal water pressure and cavity development.

Over the last ten years rapid surface melting has been documented on the Greenland Ice Cap (van den Broecke and others, 2009) that represents a modern analog for rapid delivery of meltwater to the bed of an ice cap. On the western margin of the Greenland Ice Cap large melt pools form 10-100 km (centered on 25-60 km) behind the margin of the glacier and drain, sometimes in less than two days, to the glacier bed through as much as 1000 m of ice (Zwally and others, 2002; Das and others, 2008). This rapid melting and drainage is associated with accelerated glacial flow (Zwally and others, 2002; Das and others, 2008; Joughin and others, 2008; van de Wal and others, 2008). Perhaps situations like this were occurring sporadically during the somewhat rapid recession (80-300 m/yr; Ridge, 2003, 2004, 2010) of the last ice sheet from New Hampshire and Vermont, almost certainly a time of rapid surface melting and meltwater production. Currently there are no ages for cavity deposits to determine where the margin of the glacier was when the cavity deposits were forming. However, the remanent declination of cavity deposits at Houghton Brook ($358° \pm 1.6°$) is consistent with the declination of ice proximal varves that would have formed when the receding ice sheet was 35-40 km south of Houghton Brook (Ridge, 2003, 2004, 2010). This matches the distance from the margin where melt pools form on the surface of the Greenland Ice Cap today. With more paleomagnetic measurements from subglacial cavity deposits it may be possible to test the hypothesis that cavities formed at a certain distance interval behind the receding ice front.

Conclusions

Sub-till lacustrine deposits beneath till of the last glaciation primarily represent lacustrine deposits formed in proglacial advance lakes at the front of the advancing ice sheet and also subglacial water-filled cavities. The advance lake deposits are relatively thick and rest on non-glacial surfaces. They have well sorted clay to fine sand, varves, trace fossils, and relatively low abundances of ice-rafted debris. Subglacial cavity deposits always occur within till units and are relatively thin with very abundant till pellets, dropstones, and layers of diamicton, while also showing an absence of trace fossils. Cavity deposits often have rhythmic muddy sands and clayey silts but they do not have the discrete nearly pure clay beds of varves found in proglacial lake deposits.

The preservation of sub-till lacustrine deposits from advance lakes highlights the differences between glacial erosion in east-west vs. north-south trending valley segments. Advance lake deposits in east-west valley segments document areas of no net glacial erosion while north-south valleys were more deeply eroded well below the depths of postglacial valley fill. East-west trending valleys also had a wet mobile bed when debris accumulated to produce thick till deposits. Subglacial cavity deposits document a subglacial wet-based environment with water pressure sufficient to hold open subglacial cavities beneath the last ice sheet, perhaps representing a time when surface melting of the glacier produced large volumes of water that made its way to the glacier bed.

Acknowledgments

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Field trip stop description for Jack Ridge

Sub-till Lacustrine Sediment of the Sugar River Valley

Location: Sugar River Drive on south side of Sugar River Valley in Newport, NH about 0.7 km east of the town line with Claremont.

GPS Coordinates: UTM zone 18, E721900 N4803840

This is a narrow dirt road with little room for parking. Please stay off the road and be careful!!

Description:

Sub-till Lacustrine Sediment

Beneath more than 50 ft of till at this exposure is a package of lake sediment deposited during the advance of the last glacier. The lacustrine unit is composed of varved laminated silt to fine sand and dark gray clay interbedded with diamicton beds interpreted to be debris flows. The lacustrine unit is also full of ice-rafted debris in the form of diamicton pellets (till balls) and dropstones. The geography of this valley is perfect for ice originally advancing into the valley from both ends as it squeezed around Green Mountain to the north. This may have formed a very deep lake between two advancing ice lobes that promoted calving. The lacustrine sediment extends below road level and the base of the unit is not exposed. The top of the lacustrine unit is very deformed and lacustrine sediment is incorporated into the bottom of the otherwise stony till unit above. Paleomagnetic measurements in the lacustrine sediment reveal a declination of 20-25°, which eliminates this lacustrine deposit as sediment deposited during the recession of the last glaciation when geomagnetic declination was about 320° (40°W).

From Denny (1958) Appendix A

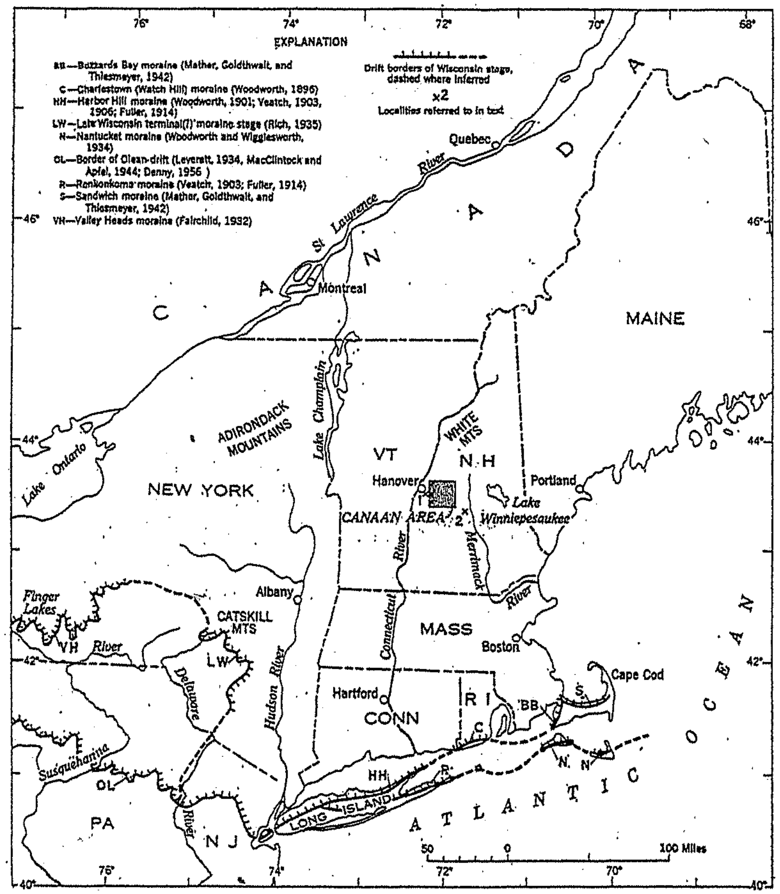


FIGURE 13.—Map of New England and vicinity showing some drift borders of the Wisconsin stage (Denny, 1956a, fig. 1C) and location of Canaan area.

weeks were spent in the area in 1954. The writer acknowledges with thanks the assistance of the society and their consent to publication outside their jurisdiction. W. H. Lyford, soil scientist in the Soil Conservation Service, U. S. Department of Agriculture, gave freely of his knowledge of the region. The work in 1940 and 1941 was carried on under the guidance of the late J. W. Goldthwait whose kindly assistance is gratefully remembered. The bedrock outcrops shown on plate 4 are based chiefly on unpublished data made available through the kindness of K. Fowler-Billings and C. A. Chapman. The author is also grateful for the assistance of S. E. White and D. C. Nutt.

A-1

PHYSICAL GEOGRAPHY

The Canaan area is in the maturely dissected uplands of central New England. Low mountains rise 1,000 to 2,000 feet above rolling hills a few hundred feet high that separate narrow flood plains, swamps, and ponds. The lowest point in the area is Mascoma Lake, altitude 751 feet; the highest Mount Cardigan, altitude 3,121 feet. Melvin Hill, the divide between Indian River and Smith River just north of Tewksbury Pond, and Mount Cardigan are on the boundary between the Connecticut and Merrimack watersheds (fig. 13). The Mascoma and Indian Rivers flow westward across the regional trend of bedrock structure for part of their courses, but most streams flow northward or southward parallel to the structure.

The region has long cold winters with heavy snowfalls and short, comparatively cool summers. The climate varies widely with altitude (U. S. Dept. Agr., 1941, p. 989–1001). Podzolic soils characterize the area (Latimer and others, 1939).

BEDROCK GEOLOGY

The Canaan area is underlain by igneous and metamorphic rocks of Paleozoic age that were deformed and metamorphosed during late Paleozoic time (pl. 5). Most of the rocks are foliated. The foliation strikes northward.

The bedrock influences the location of major topographic features. Quartz-rich rocks in the Clough quartzite (pl. 5) underlie highlands along the western edge of the area. To the east, foliated igneous rocks, including granite, quartz monzonite, and granodiorite, and the Bethlehem gneiss form lowlands and rolling hills. The divide between the Merrimack and Connecticut Rivers follows a highland where the Littleton formation is cut by many pegmatites close to its contact with Bethlehem gneiss. The summit of Mount Cardigan is composed of Kinsman quartz monzonite, but its western slope and much of the divide to the north is underlain by quartzite and sillimanite schist in the Littleton formation. In summary, the Clough quartzite and Littleton formation, rocks with abundant pegmatite, and the Kinsman quartz monzonite underlie highlands; foliated igneous rocks and the Bethlehem gneiss underlie lowlands.

SURFICIAL GEOLOGY

The surficial mantle includes till and kame deposits of Wisconsin age and alluvium and swamp deposits of Recent age. The kame deposits are well exposed in numerous borrow pits, but the other surficial deposits are largely concealed beneath the vegetation. Contacts on plate 4 are drawn primarily on topographic evidence; rarely is a contact exposed. Contacts are dashed lines where topographic evidence is equivocal or where designation of a surficial map unit is queried.

WISCONSIN DRIFT

TILL

Two varieties of till are distinguished in the Canaan region. One is a loose, bouldery material with a coarse sandy matrix. The other is a compact till containing few boulders. Commonly the loose till rests on the compact till. This distinction between a loose upper till and a compact lower till was made originally in New Hampshire by Upham who later adopted Chamberlin's terms "englacial" and "subglacial" drift (Torell, 1877; Upham, 1878, p. 9–10; 1891; Crosby, 1890; Chamberlin, 1894; Flint, 1947; Lawrence Goldthwait, 1948). Both varieties of till are found throughout the area, but loose till is extensive at the surface where the bedrock is coarse grained and massive, and compact till is widespread where the bedrock is fine grained and schistose. On plate 4, the till is not divided into the two varieties. The loose, bouldery till was probably deposited in part directly by ice and in part by water. The compact till was deposited beneath moving ice. Some geologists have interpreted a loose till resting on compact till as the deposits of two glaciations (Currier, 1941; Moss, 1943; White, 1947; Judson, 1949). Some data on the physical properties of till in New Hampshire are given by Lawrence Goldthwait (1948).

COMPACT TILL

This till is olive-gray to grayish-brown and its matrix ranges in texture from fine sand to silt loam (Soil Survey Staff, 1951, p. 205–213). Most till with a silt loam matrix is found where the adjacent bedrock is the Littleton formation or the Ammonoosuc volcanics (pl. 5). Compact till contains from 5 to 10 percent pebbles, cobbles, and boulders. Many rock fragments are abraded and striated, especially those that are fine grained. Most compact till is a heterogeneous mass of material, but in some exposures, there are faint wavy laminae of fine sand. Compact till commonly has a platy structure, the plates being parallel to the ground surface. The plates increase in size from about 1 inch long near the surface to lengths of from 3 to 6 inches and thicknesses of from 1 to 2 inches at depths of 10 feet or more. In the few deep excavations studied, platy structure was not visible in compact till at depths of more than 10 or 15 feet (tables 1 and 2).

A-Z

TABLE 1.—*Section of compact till in area of schistose bedrock. Exposure in east bank of Stony Brook about 1 mile south of State Route 14, or about 3 miles west of Enfield (loc. 1, fig. 13; outside of area shown in pl. 4)*

	Approximate thickness (feet)
Top.	
1. Sand; bouldery, yellowish-brown, noncalcareous, loose, unstratified	3
Contact sharp.	
2. Till; silt loam matrix; dominantly dark brownish-gray; locally mottled with brown; noncalcareous; compact; platy structure; contains pebbles, cobbles, and a few boulders, all with a yellowish-brown silty coating. Some pebbles of gneiss are soft; others are hard	9
Contact gradational through about 6 inches.	
3. Till; silt loam matrix; mottled olive-gray and grayish-brown; dominantly olive-gray near base; calcareous, firm; compact; structureless	6
4. Till; similar to that above except transititional in color from that above to that below	3
5. Till; silt loam matrix; olive-gray; calcareous; firm; compact; structureless; contains perhaps a maximum of 10 percent of pebbles, cobbles, and boulders, many striated. About half of the fragments are of relatively fine-grained schistose rock. In places contains a few layers of sand as much as a quarter of an inch thick	48
Base not exposed.	

TABLE 2.—*Section of compact till exposed in borrow pit on north side of State Route 11 about 1 mile west of Webster Lake, Penacook quadrangle, New Hampshire (loc. 2, fig. 13; about 19 miles southeast of Canaan)* [1]

	Approximate thickness (feet)
Top.	
1. Outwash or till; boulders, cobbles, and pebbles in a coarse sandy matrix; yellowish-brown; loose; unstratified	10
Contact sharp, wavy.	
2. Till; sandy loam matrix; dark grayish-brown (2.5Y 4/2); noncalcareous (pH 7.0–8.0); hard; compact; platy structure; contains about 5 percent of rock fragments, most less than a half inch in diameter. Plates are curved and near base of unit 2 are approximately 6 inches by 3 inches by 1 inch; plates are progressively smaller toward top. Surface of plates covered by single layer of sand grains. Some fragments of schistose rock can be crushed between the fingers	10
Contact gradational through about 4 inches.	
3. Till; loam matrix; olive-gray to dark olive-gray (5Y 4/2–3/2); very slightly calcareous: hard; compact; structureless; contains a few percent of fragments of schistose and coarse-grained igneous rocks. A very few fragments are stained brown and can be crushed between the fingers	40
Base not exposed.	

[1] Density, moisture content, and mechanical analyses of samples of till from this pit are given by Lawrence Goldthwait (1948). Locality listed as "Andover, on Rt. 11 'Borrow Area A' for Franklin Dam."

The compact till occurs in all topographic positions but is probably most extensive on lower slopes, especially on north-facing and north-west-facing hillsides. At or near summits the till is thin or bedrock is exposed. The maximum exposed thickness of compact till is about 30 feet, but thicknesses of more than 50 feet are readily inferred. For example, Gulf Brook flows northward from Spectacle Pond in Enfield in a gorge nearly 80 feet deep apparently cut in compact till (pl. 4).

Along the edge of valley floors compact till is overlain by younger deposits. Its extent or thickness beneath such deposits is unknown. On some slopes and on most hilltops compact till abuts against rock outcrops with no topographic break. In other places a gentle slope on compact till near a valley bottom extends uphill to a steep slope on bedrock thinly veneered with drift.

LOOSE TILL

The loose till is an olive-gray to white mixture of sand and silt with little clay (texture, coarse sand to loam). Boulders, cobbles, and pebbles are perhaps as much as 25 percent of the total mass. The rock fragments are angular and subangular; striated fragments are scarce. Most loose till is irregularly and discontinuously stratified, the beds lensing out within distances of a foot or two (figs. 14 and 15). A characteristic feature is small curving lenses of sand ranging from 2 inches to 1 foot long and from ¼ inch to 3 inches thick in an otherwise massive deposit. In places a pebble-free deposit consists of thin, curved layers of fine sand that resemble lake beds.

Two sketches illustrate the structure of loose till. In figure 14, unit 1 is a bouldery unsorted mixture of sand containing a few curved laminae of silt. The unit could have been deposited directly from melting ice. However, units 2 and 3 are more or less stratified and seem to require moving water for their deposition. The layers of unit 2 curve around the base of the overlying boulders and in unit 3 the stratification is discontinuous, curved, or inclined. The underlying compact till, unit 4, is firm and compact in comparison with the overlying material and contains relatively few fragments of rock.

In figure 15 none of the units are typical till. Unit 7 resembles lake beds and is intertongued with the coarse, almost structureless sand of unit 6. Unit 4 appears to be water-laid gravel and sand. Unit 3 is massive coarse sand containing fragments of laminated fine sand. Unit 2 is partly massive, partly stratified, and contains only a few pebbles.

Near South Grafton School and elsewhere a very sandy loose till

A-3

and topographically resemble kame deposits (fig. 16). There is a gradation in lithologic character, in structure, and in topographic expression from loose till into kame deposits. On plate 4, kame deposits are restricted to well-sorted and well-stratified materials forming distinct knolls, ridges, or terraces.

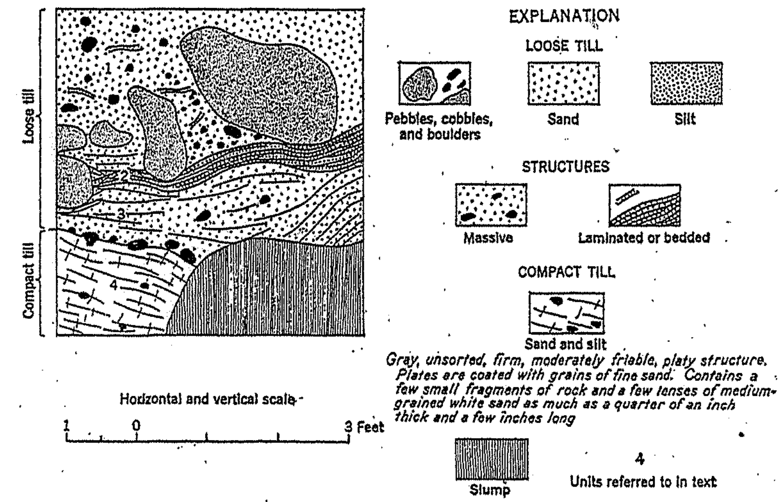
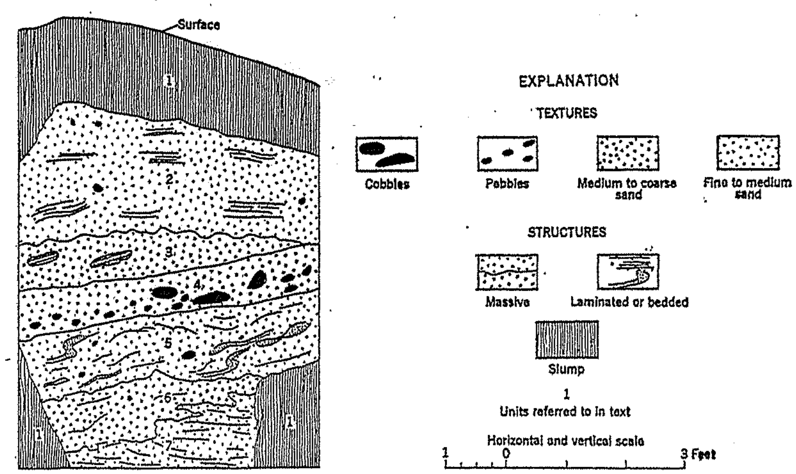


FIGURE 14.—Loose till. Unsorted mixture of pebbles, cobbles, boulders, and sand overlying laminated sand. Rests on compact till. Top of figure about 3 feet below ground surface. Exposure in road cut, altitude 1,020 feet, on west side of unsurfaced road about 1.2 miles south of Tuttle Hill in Orange (loc. 4, pl. 4).



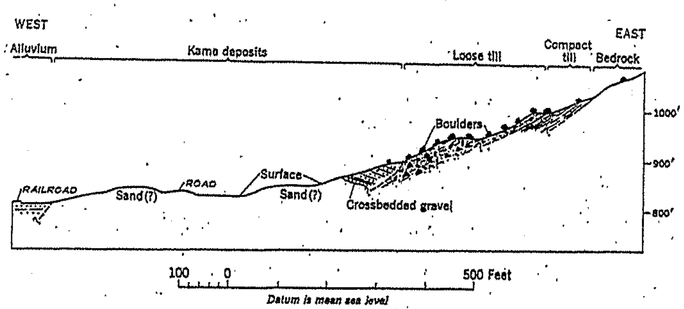


FIGURE 16.—Boulder-covered knolls of loose till on lower part of valley wall just above top of adjacent kame deposits. Locality is the southwest slope of Pine Hill on east side of Smith River in Danbury (loc. 7, pl. 4). Cross section based on a field sketch and plate 4.

Loose till is best exposed and probably most extensive on lower slopes just above the top of the kame deposits. Sections as much as 30 feet thick are exposed. Loose till appears to be absent from large areas on the uplands.

STRATIGRAPHIC RELATIONS

Loose bouldery till resting on compact till was seen at several exposures. In figure 17 a boulder-strewn knoll is underlain by about 8 feet of loose till that grades through a zone about 3 inches thick into firm, compact till with platy structure. Loose till was nowhere seen to overlie bedrock, but compact till has been seen to rest on bedrock at a number of places. No loose till has been found beneath compact till.

ORIGIN

The compact till and the loose till were deposited by one ice sheet but are unlike because of differences in manner of deposition. The field evidence demonstrates that loose till is bouldery and sandy, that compact till is loamy and includes fewer boulders than loose till, that loose till overlies compact till, that small masses of one variety of till occur within the other type, and that much loose till underlies knolls on lower valley walls. Loose till is less extensive on uplands and mountains.

These field relations are for the most part those observed by Upham (1878) but, in addition, he noted that most fragments of rock in the lower till were abraded to a considerable extent and some came from distant sources, whereas the fragments in the upper till were less abraded and locally derived, perhaps from adjacent hilltops. Upham concluded (1878) that the upper till was debris carried on top of or within the ice sheet, whereas the lower till was carried and deposited

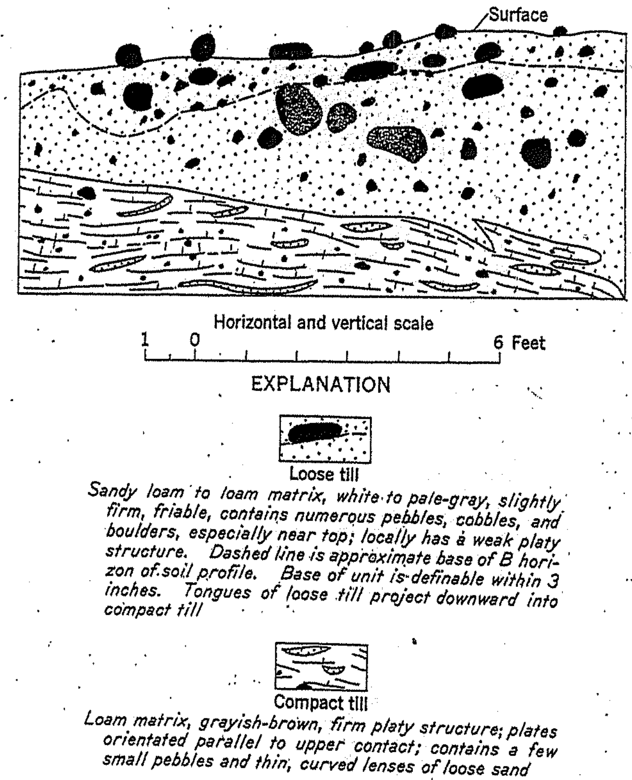


FIGURE 17.—Loose bouldery till resting on compact till. Exposure along dirt road at corner, altitude 1,220 feet, about half a mile west of Shepard Hill in Grafton (loc. 5, pl. 4).

under the ice. Upham's use of "upper till" and "lower till" does not entirely correspond to the writer's use of "compact till" and "loose till." Upham implied that the upper till formed an almost continuous surface mantle. He included in the upper till the soils developed on compact till.

The two varieties of till probably originated in the following way: Compact till was deposited beneath the glacier and doubtless owes its compactness partly to the weight of the overlying ice. Loose till, on the other hand, accumulated beneath, within, or on top of the ice, probably when it was melting upward from the bottom as well as downward from the top. Water-laid drift was deposited beneath or within the ice and subsequently was overlain by unsorted debris from around the lower part of the boulders (fig. 14) suggest that the boulders were let down from ice above after the laminae of sand and silt were deposited.

Boulder-strewn knolls of loose till resting on compact till and in turn overlain by kame deposits (fig. 16) suggest the following sequence of events: As the marginal zone of the glacier wasted downward, bouldery loose till was dropped on compact till, at first from beneath the ice with some water-laid material, later by sliding from the surface of the ice or from the adjacent valley wall. Streams flowing in channels between wasting ice and the valley wall laid down kame deposits on or against the loose till. Boulders were dropped or slid from the ice or from the valley wall on to both till and kame deposits. The relative scarcity of boulders within some loose till and most kame deposits compared with their abundance on the surface suggests that such bouldery accumulations result from surface sliding rather than as a lag concentrate.

The platyness of the compact till is commonly regarded as a structure resulting from compression of the till caused by the weight of the overlying ice (Flint, 1947, p. 106–107). However, this platyness may be primarily a result of weathering (p. 87).

GLACIOFLUVIAL SEDIMENTS

The glaciofluvial deposits are chiefly sand, but about one-fifth are gravel. These materials were mapped as kame deposits and terrace sand (pl. 4). The sand is both coarse and fine grained, light gray to white, commonly includes scattered pebbles, and much of it is well sorted. The gravel consists chiefly of pebbles and a few cobbles in a coarse sandy matrix. Boulders are rare. The glaciofluvial deposits form horizontal strata ranging from 1 inch to 2 feet thick; a common thickness is 2 or 3 inches. Crossbedding (fig. 18) is a common feature of both the sand and the pebble gravel. Within one borrow pit the foresets may dip in several directions. The glaciofluvial deposits are known to rest on till or bedrock. The terrace sand is assumed on topographic evidence to rest on or against the adjacent kame deposits, but the contact was not exposed.

KAME DEPOSITS

The water-laid drift mapped as kame deposits forms terraces or ridges that extend northward along lower valley walls or valley floors. The deposits are more extensive in northward-draining valleys than in those that drain south and are a little more numerous on the east side of valleys than on their western slopes. Sandy kames contain less than about 40 percent gravel, commonly not more than 10 percent. They are more abundant than the gravelly kames. Kame deposits locally resemble loose till. On plate 4 kame deposits

A-5

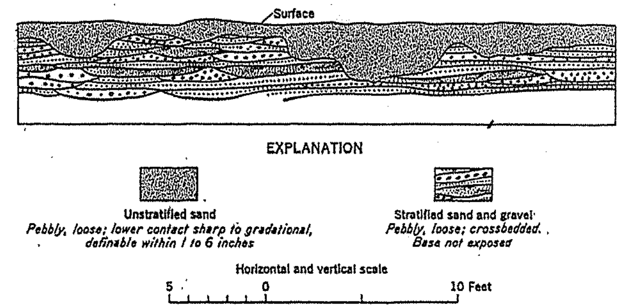


FIGURE 18.—Crossbedded kame deposits along Smith River in Danbury. Upper part of deposit is unstratified, the result of mechanical disturbance caused chiefly by roots of trees. Exposure in borrow pit on east side of U. S. Route 4, 1.4 miles northwest of Danbury (loc. 8, pl. 4).

are shown only where the materials are well sorted, well stratified, and form conspicuous knolls, ridges, or terraces. A few poorly exposed masses, apparently water-laid drift, are mapped as probable kames (Qk?).

Many sandy kames consist of 1 to 2 feet of crossbedded pebble gravel overlying stratified sand. The sand commonly contains lenses of crossbedded gravel and is probably a stream deposit; such kames are those in the valley of Little Brook south of George Pond and those that extend southeastward from Mirror Lake into and down the valley of Smith River to Danbury. In some sandy kames, the sand forms regular beds ranging from a fraction of an inch to about 2 inches in thickness and showing small-scale crossbedding or ripple marks. Such sand, probably of lacustrine origin, is exposed in borrow pits west of Mirror Lake and west of Melvin Hill (fig. 19). In the valley of Haines Brook south of Canaan, and in a small valley just west of Danbury, sandy kame deposits, at least in part of lacustrine origin, are covered with boulders as much as 10 feet in diameter, a few of which have been sand-blasted, especially those in the southern part of each valley. In places the sand is horizontally stratified; elsewhere foreset(?) beds dip southward.

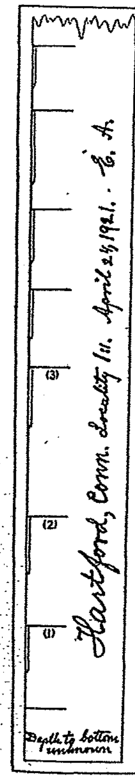
The gravelly kames are small deposits of gravel that commonly form narrow ridges. A string of such ridges east of the junction of Indian River and Mascoma River was probably deposited in a channel in the ice. Exposures show southward-dipping beds.

Most kame deposits are at least 20 feet thick. The greatest thickness exposed or readily inferred is about 80 feet.

Near the margins of many kame deposits, the beds have been displaced along both normal and reverse faults with displacements of as much as several feet. Many of the fault planes strike parallel to

B-1

—View showing the method of measuring the clay layers on a strip of paper. lty 67, Hanover, N. H.



to measure in the field, it is advisable to take a series of samples to measure at home. Samples of different types of clay also make valuable material for museum exhibit and teaching. The samples are taken in tight troughs of zinc plate, 50 centimeters long, 5 centimeters wide, and 2 centimeters high. The face of the clay bank is carefully smoothed, and the trough is cautiously pressed in, a knife being used to cut away the clay just outside the edges, until the trough is entirely filled. The samples are taken so that they overlap each other 5 centimeters. The troughs are then cut loose from the bank, and part of the clay which projects above the edges is removed. If moistened and wrapped in wax paper, such samples keep in good condition for months. In the laboratory the clay is carefully cut

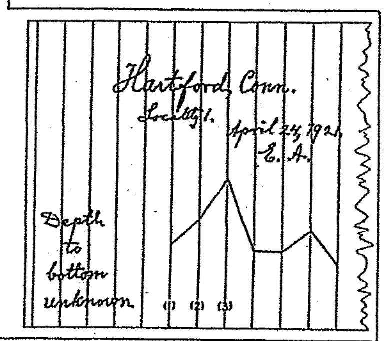


FIG. 4—Sample of measurement made in the field and curve constructed from it in the office. Actual scale.

From Antevs (1922) - Appendix B

LOCALITIES IN THE CONNECTICUT VALLEY (*Continued*)

61.—5 miles S of Windsor, Vt., bluff on the brook S of Ascutneyville, between the highway and the Connecticut River.

Plain of sedimentation.
4 feet gravel.
2 feet sand and silt.
Varve sediments, silt in the upper half, sand in the lower half.
25 feet glacial gravel.
Bed rock.

Series measured: 6601–6662, (6608–6638).

In the same bluff, a little to the east, the section is essentially different because of disturbances and downslidden horizons.

Number 6601 is bottom varve.

62.—3 miles NW of Claremont, N. H., bluff on the highway, just W of the junction of the Sugar River, the railroad, and the highway.

The material is silty. The thickness of the varves varies from 6 inches to 2½ feet. Depth of the deposit is unknown.

Series measured: (6613–6631).

63.—3½ miles NW of Claremont, ¾ mile N of the railroad bridge across the Sugar River, where the railroad crosses Walter Brook.

Several measurements were carried out in slides in the deep brook ravine as far as ½ mile NNE of the railroad as well as in the railroad banks N and S of the brook. 600 yards NNE of the railroad the profile was like this:

Plain of sedimentation.
15 feet sand.
Varve silt to sand with thin winter layers, varves 6806–7073.
Varve clay to silt with very good lamination, varves 6642–6805.
Varve silt, somewhat sandy, with several slidden zones. Thickness of varves from 2 inches to 2 feet. About 30 varves measured but not connected with the normal curve.
6 feet talus down to the brook. Depth to substratum probably inconsiderable.

Series measured: 6642–6662, 6643–6666, 6647–6713, 6648–6658, 6680–6706, 6690–6805, 6714–6823, 6726–6823, 6811–6823, 6858–6981, 6858–6922, 6858–6898, 6858–6871, 6894–7073, 6930–6980.

64.—1 mile SSE of Westboro, N. H., ravine on the southern side of the Mascoma River, S of the first railroad bridge across the river.

Plain of sedimentation.
4 feet fine sand.
Typical varve clay.
80 to 100 feet till down to the river.

Series measured: 6750–6802, (6734, bottom varve–6749).

65.—1 mile SE of Westboro, cut on the road 150 yards NE of the new bridge across the Mascoma River and the railroad.

The clay shows good lamination. Depth to bottom is unknown.

Series measured: (6760–6801).

66.—¾ mile N of White River Junction, Vt., ravine just W of the railroad.

The clay shows rather good lamination. It is covered by 4 feet of sand. Below the measured series the varves, which are slidden, become thick and silty. Depth to substratum is more than 30 feet.

Series measured: 6783–6888.

67.—¾ mile SSE of Hanover, N. H., bluff on Mink Brook.

25 feet sand and silt. Varve limits hardly distinguishable.
Varves 6806–7040, silty, often sandy and with lenses of sand. Varve limits often very difficult to distinguish. Thicknesses varying greatly, but on the average about 2 inches.
Varves 6760–6805, silty, distinct. Thickness varying from 1 to 15 inches.
Exposed to the level of the brook. Depth to bottom unknown.

Series measured: (about 6760 to about 7040; only varves 6806–6862 have been used for control of the number).

B-1

B-2

B-2

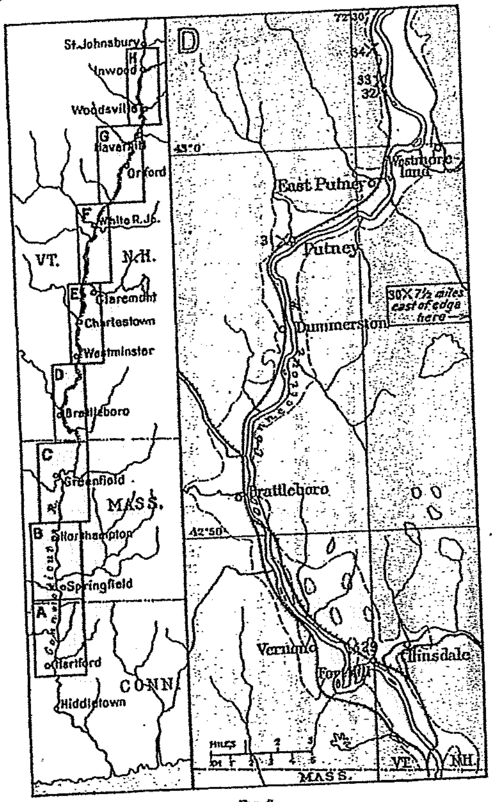


FIG. 8

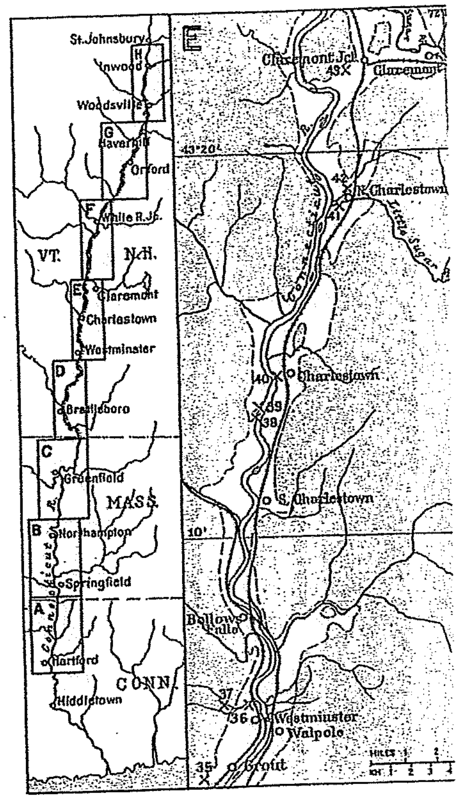


FIG. 9

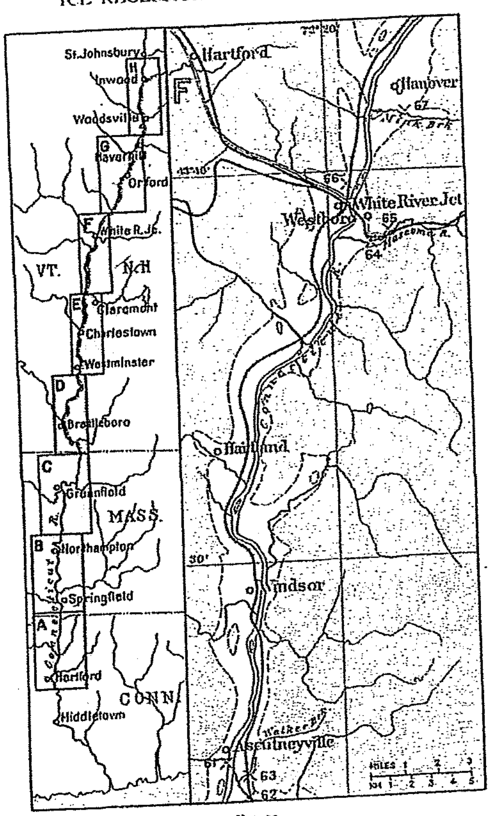


FIG. 10

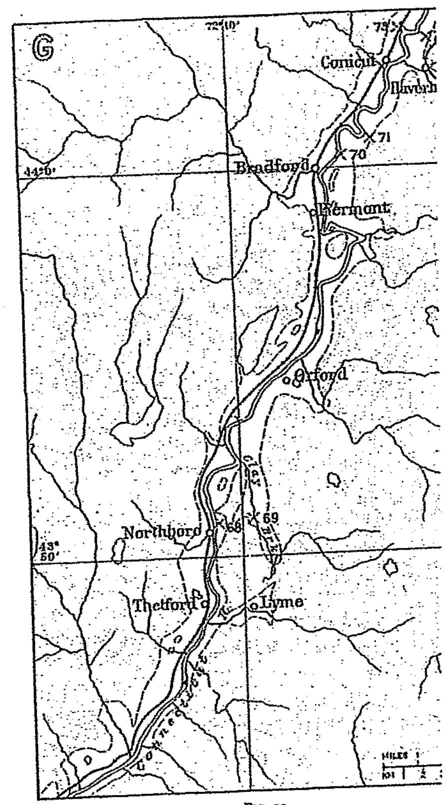


FIG. 11

B-3

CHAPTER VII

THE RATE OF RECESSION AND CONDITIONS CONTROLLING RECESSION

On account of the very considerable depth of the varve sediments in the New England valleys their bottoms have been reached only at a limited number of localities, almost all of which are situated in New Hampshire and Vermont. Here the rate of retreat, accordingly, has been exactly determined in several instances, while that has not been possible in Connecticut and Massachusetts. However, the bottom of the clay has in many cases been almost reached, enabling an approximate determination of the speed of the retreat.

On the map, Plate VI, the positions of the ice edge have been marked for every 100 years. Which of these positions are determined by bottom varves may be deduced from their location with reference to the localities indicated in Chapter III at which bottom varves were reached. Accelerated retreat between two points indicated by the speed south and north of them has been shown on the map by putting the 100-year lines at different distances, so as to illustrate as truly as possible the gradual increase. Retardation in the recession has been marked in a corresponding way.

The extent of the examined field, which reaches from Hartford, Conn., to St. Johnsbury, Vt., is 185 miles (298 kilometers). The retreat across this belt is registered by a continuous series of annual clay layers except for a narrow gap at Claremont, N. H., probably representing 200 to 300 years. The time occupied by the recession from Hartford to St. Johnsbury was about 4,100 years. This makes an average rate of 22 years to a mile, or of 238 feet (73 meters) a year.

The following table gives the rate of retreat of the ice border

between the points indicated. The distances are mea[sured] parallel to the direction of the ice movement.

TABLE I—RATE OF RETREAT OF THE ICE BORDER

Localities	Distance	Time of Retreat in Years	Rate [of] Retre[at a] Ye[ar]
1 to 7 None with bottom	24.2 miles = 127,776 feet 38.9 km.	About 525	243 f 74 r
7 to 17 None with bottom	5.2 miles = 27,456 feet 8.4 km.	About 330	83 fe 26 m
7 to 24 24 with bottom	19 miles = 100,320 feet 31 km.	>1,000	<100 f <31 r
7 to 28 None with bottom	35.8 miles = 189,024 feet 57.6 km.	About 1,630	116 f. 35 n
24 to 28	16.8 miles = 88,704 feet 27.1 km.	About 270	328 fe 100 n
28 to 29 None with bottom	12.8 miles = 67,584 feet 20.8 km.	About 350	193 fe 59 m
29 to 31 None with bottom	12.4 miles = 65,472 feet 20.1 km.	About 275	238 fe 73 m
31 to 40 40 with bottom	19.5 miles = 102,960 feet 31.4 km.	>300	<343 fe <105 m
45 to 59 Bottom almost reached at both	15.2 miles = 80,256 feet 24.4 km.	About 275	292 fe 89 m.
59 to 60 Depth to bottom at 60 unknown	7.2 miles = 38,016 feet 11.6 km.	<173	>220 fee >67 m.

TABLE I—RATE OF RETREAT OF THE ICE BORDER—*Continued*

Localities	Distance	Time of Retreat in Years	Rate of Retreat a Year
61 to 64 Both with bottom	15.5 miles = 81,840 feet 25 km.	133	615 feet 188 m.
64 to 67 67 practically with bottom	4 miles = 21,120 feet 6.4 km.	<26	>812 feet >246 m.
67 to 70 Both practically with bottom	21 miles = 110,880 feet 33.8 km.	>141	<786 feet <240 m.
70 to 71 71 with bottom	½ mile = 2,640 feet 0.8 km.	3	880 feet 268 m.
70 to 75 75 with bottom	6 miles = 31,680 feet 9.7 km.	29	1092 feet 333 m.
75 to 79 Both with bottom	5 miles = 26,400 feet 8 km.	24	1100 feet 335 m.
79 to 86 86 practically with bottom	11 miles = 58,080 feet 17.7 km.	About 70	830 feet 253 m.

RATE OF RECESSION IN THE SOUTHERN ZONE

In the southernmost zone, between Hartford and Springfield, the recession was consequently rather fast, amounting to about 243 feet (74 m.) a year. In Massachusetts the rate decreased considerably, averaging between Springfield and Greenfield only 116 feet (35 m.) annually. The recession, however, seems to have varied greatly in speed in different zones and even to have been interrupted by readvances. Between localities 7 and 24 the retreat was less than 100 feet (31 m.) a year.

PROBABLE OSCILLATIONS OF THE ICE BORDER IN THE AMHERST-NORTHAMPTON REGION

In November, 1921, there was, at locality 24, southeast of Amherst, an excellent long vertical section, part of which is shown in Figure 15. The sequence of strata is clear on the left side of the section. The thicknesses are as follows, beginning at the top:

2½ feet till.
2½ feet crumpled clay, about 150 varves.
10 feet excellent clay, varves 4450–4668.
1¾ feet till.
More than 1¼ feet quicksand.

To the right, the whole clay bed is crumbled down to the till. The pressure which folded the clay seems to have come from the west-northwest. The connection between the covering til[l] and the folding of the clay is evident, but the circumstances under which the till was deposited are not clear. Those smal[l]

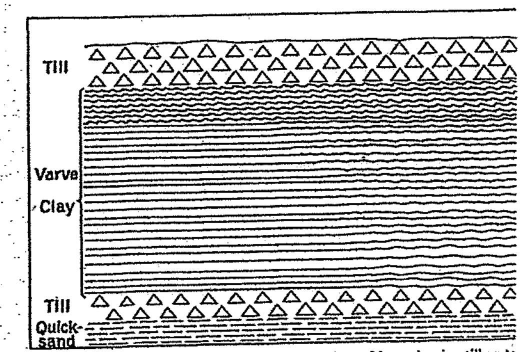


FIG. 15—Section at locality 24, southeast of Amherst, Mass., showing till on to[p] of the varve clay and partial folding of the clay by pressure from the west-northwes[t]

Appendix D D-1

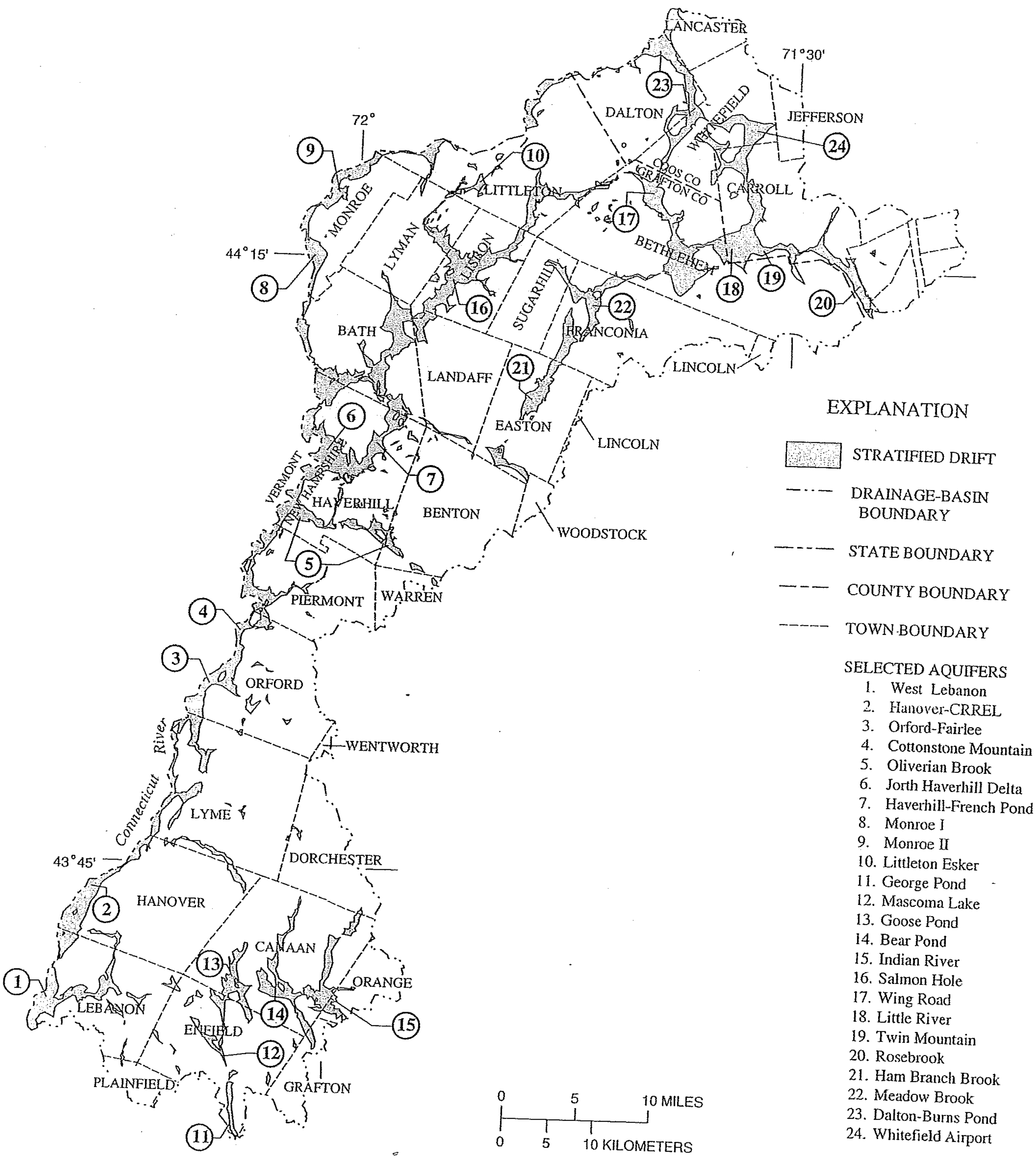
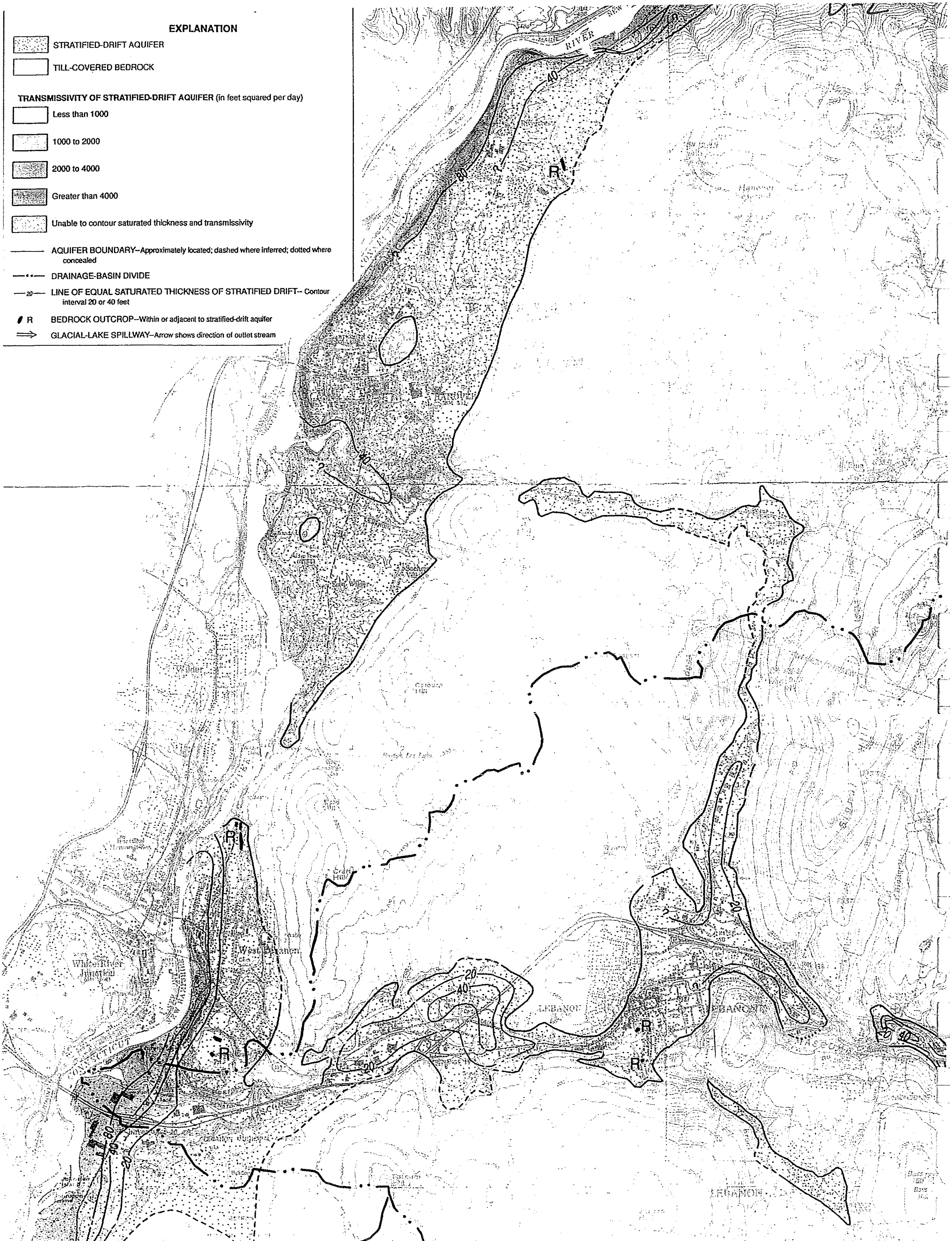


Figure 11. General locations of selected stratified-drift aquifers in the Middle Connecticut River Basin, west-central New Hampshire.

From Flanagan (1996)


EXPLANATION
STRATIFIED-DRIFT AQUIFER
TILL-COVERED BEDROCK
TRANSMISSIVITY OF STRATIFIED-DRIFT AQUIFER (in feet squared per day)
Less than 1000
1000 to 2000
2000 to 4000
Greater than 4000
Unable to contour saturated thickness and transmissivity
AQUIFER BOUNDARY–Approximately located; dashed where inferred; dotted where concealed
DRAINAGE-BASIN DIVIDE
LINE OF EQUAL SATURATED THICKNESS OF STRATIFIED DRIFT– Contour interval 20 or 40 feet
R BEDROCK OUTCROP–Within or adjacent to stratified-drift aquifer
GLACIAL-LAKE SPILLWAY–Arrow shows direction of outlet stream
RIVER
40
80
R
Wilder
West Lebanon
White River Junction
CONNECTICUT
LEBANON
20
40

D-3

Description of Selected Stratified-Drift Aquifers

Stratified-drift aquifers found in valleys throughout the Middle Connecticut River Basin underlie 123 mi^2, or 12.5 percent of the study area. The most extensive and productive (or potentially productive) aquifers are described in this section (fig. 11). Each area is referred to by the following major units: the Connecticut River Valley, Mascoma River subbasin, Ammonoosuc River subbasin, and John's River subbasin. Underlying aquifers present in each area are discussed from south to north or west to east. Aquifer boundaries, data-collection locations, and altitudes of ground-water tables are shown on plates 1 through 4. Aquifer boundaries, saturated thickness, and transmissivity of stratified-drift aquifers are shown on plates 5 through 8. Brief discussions of information shown on the plates are included in the following section.

Connecticut River Valley Aquifers

Major stratified-drift aquifers along the main stem of the Connecticut River in the study area extend north from Lebanon to Littleton and underlie 43 mi^2. Other towns with underlying stratified-drift aquifers in the Connecticut River Valley are, from south to north, Hanover, Lyme, Orford, Piermont, Haverhill, Bath, and Monroe (fig. 11). Aquifers underlying areas drained by Oliverian Brook, Clark Brook, and other minor tributary streams also are included in this section.

Connecticut Valley Esker Aquifers

Hitchcock (1878) was the first to describe a long and narrow, south-trending esker ridge, known as the Connecticut River Valley esker, that formed beneath glacial ice along the floor of the Connecticut River Valley. Hitchcock (1878) and Lougee (1939) believed that the esker formed as a 24-mile-long continuous deposit from Windsor, Vt., to Lyme, N.H., and that much of the esker is either buried beneath fine-grained lake-bottom deposits or partly eroded away by the Connecticut River. More recently, Koteff and Pessl (1981) have postulated that this esker formed as part of the systematic process of stagnation-zone retreat. Two aquifers (one in Lebanon and the other in Hanover) that are part of the Connecticut River Valley esker are described below.

D-4

West Lebanon Aquifer

The southernmost aquifer originally deposited in glacial Lake Hitchcock along the main stem of the Connecticut River as part of the Connecticut River Valley esker is west of New Hampshire Route 12A in the southwestern corner of Lebanon (fig. 11, pl. 5). Much of this stratified drift consists of fine-grained lake-bottom deposits either overlain by, or interlayered with, coarse-grained fluvial-deltaic material. Coarse-grained aquifer material underlies the northern part of the aquifer and may be part of the Connecticut River Valley esker that is buried beneath younger lake-bottom deposits. Bridge boring LHB-1 penetrated 65 ft of saturated coarse sand and gravel, overlain by 15 ft of fine sand and silt; the boring did not reach bedrock (pl. 1, appendix B). The esker continues north of boring LHB-1 and is exposed at the surface near the confluence of the Mascoma and Connecticut Rivers. Bridge boring LHB-4 penetrated the esker segment 6,000 ft northeast of boring LHB-1. The stratigraphic log from boring LHB-4 (appendix B) shows that the esker contains 33 ft of medium sand and gravel overlain by 29 ft of fine to medium sand; here also, the boring did not reach bedrock. Surface features shown on the Hanover, N.H.-Vt., USGS 7.5 by 7.5 minute quadrangle map (pl. 1) indicate that the esker continues south of boring LHB-1 across the Conneccut River into Hartford, Vt.

Saturated thickness in the West Lebanon aquifer exceeds 80 ft in some sections. Transmissivity generally is less than 4,000 ft^2/d. Potential for ground-water production is considerable, especially if production were close to the Connecticut River where recharge can be induced from the river. However, because of the proximity of the aquifer to the town landfill, the town wastewater treatment plant (at the confluence of the Mascoma and Connecticut Rivers), and urbanization, water quality may be of concern.

Hanover-CRREL Aquifer

This aquifer is located in western Hanover, immediately west of the U.S. Army Cold Regions Research and Engineering Laboratory (CRREL) (fig. 11, pl. 5). The aquifer is hydraulically connected with the Connecticut River. Near CRREL, the esker is overlain by approximately 90 ft of silt and clay lake-bottom sediments. Some of the wells drilled into the aquifer penetrated the full thickness of the esker and reached bedrock. At observation well HHW-28, bedrock is 65 ft below land surface. Industrial well HHW-5 penetrated 88 ft of saturated coarse-grained material and refusal was reached on bedrock at 167 ft. Four other industrial wells (HHW-2, HHW-3, HHW-4, and HHW-6) and Hanover Water Works municipal well HHW-1 (located 1,000 ft north of the industrial wells) penetrated the esker. Although the esker segment is long (at least 4 mi), it is very narrow, probably 1,000 ft wide or less. The contact between the esker and surrounding lake sediments is fairly abrupt and steep. Test well HHW-27, which penetrated 139 ft of fine sands, silt, and clay, is less than 600 ft east of industrial well HHW-4, which penetrated 42 ft of coarse-grained aquifer material (appendix B).

Information on the hydraulic characteristics of the aquifer is available from a study of the Norwich, Vt., Fire District municipal well NRW-42, 4,300 ft northeast of well HHW-1 on the opposite side of the Connecticut River (pl. 1). Well NRW-42 yields nearly 1,000 gal/min from this highly permeable esker, and similar hydraulic conditions are at the industrial wells and Hanover municipal well (Shoop and Gatto, 1992). Based on results from aquifer tests on well NRW-42, transmissivity is 36,800 ft^2/d and average hydraulic conductivity is 360 ft/d (table 3) (Caswell, 1990). This transmissivity, however, represents the effect of a recharge boundary (the Connecticut River), which increases availability of water to the aquifer; hence, this transmissivity is not representative of the hydraulic characteristics of the aquifer.

Currently (1993), water is withdrawn from this aquifer at a rate of 2.9 to 3.5 Mgal/d. CRREL uses the aquifer as a source of industrial water for its refrigeration system; the combined yield for CRREL's four industrial wells is 1,500 to 1,900 gal/min (2.2 to 2.8 Mgal/d). Hanover Water Works used to pump from the aquifer during the summer months to augment the town's three surface-water reservoirs; well HHW-1 had a yield of 500 gal/min (0.63 Mgal/d). The Norwich (Vt.) Fire and Water District withdraws from the aquifer as its primary source of drinking water. Although the District's well (NRW-42) is capable of yields of 725 gal/min or 1.04 Mgal/d (Caswell, 1990),

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it currently (1993) yields only 70 gal/min or 0.1 Mgal/d (Brian McMullen, Norwich Fire and Water District, oral commun., 1993).

Ground-water levels reported for observation wells HHW-25, HHW-26, and HHW-27 (appendix A) reflect nearby pumping conditions from the industrial wells. These water levels do not match the altitudes of the water table shown on plate 1, which are estimates of natural or nonpumping conditions. The actual direction of ground-water flow—not shown on plate 1—is from the Connecticut River to the industrial wells, in response to induced infiltration from the river.

Additional water-supply wells could be developed south or north of the current wells; however, trichloroethylene (TCE) was detected in November 1990 in three of the CRREL industrial wells screened in the stratified-drift aquifers and in the underlying bedrock (Shoop and Gatto, 1992). Hanover well HHW-1, used only intermittently for drinking water, has not been contaminated with TCE; however, it was shut down in 1991 because of its proximity to the contaminated wells. Water from well HHW-1 also contained high concentrations of iron and did not mix well with chlorinated surface water (Ed Brown, Hanover Water Works, oral commun., 1993). Water from the industrial wells continues to be used for cooling purposes only. The Norwich (Vt.) Fire District municipal well is sampled regularly for water quality and is presently (1993) still in use.

Orford-Fairlee Aquifer

The Orford-Fairlee aquifer is in the western part of Orford just north of Reeds Marsh (fig. 11, pl. 6). The most productive part of the aquifer is where Jacob's Brook once drained into glacial Lake Hitchcock. Here, coarse-grained deltaic deposits overlie fine-grained lake-bottom deposits. The aquifer is hydraulically connected to the Connecticut River. Seismic-refraction profiles (Orford a-a', pl. 2, appendix C.29) show the saturated thickness to be 200 to 340 ft at the southern end of the aquifer near Reeds Marsh and 170 to 210 ft at the confluence of Jacob's Brook and the Connecticut River (Orford e-e', pl. 2, appendix C.31). Orford domestic well OSW-47 (pl. 2), at the southern end of seismic-refraction line Orford e-e', penetrated 83 ft of sand; the well did not reach bedrock (appendix B). USGS observation well OSW-2 (pl. 2), at the western end of seismic-refraction line Orford a-a', penetrated alternating layers of coarse-grained aquifer material and fine-grained lake-bottom deposits (fig. 12) (appendix B).

Although the aquifer remains undeveloped on the New Hampshire side, it presently provides municipal water across the State border in Fairlee, Vt. Across the river from Orford in Fairlee, Vermont (pl. 2), water is withdrawn from municipal well FLW-1 at a rate of only 70 gal/min (0.1 Mgal/d), but it is capable of yielding as much as 1.0 Mgal/d (Lance Colby, Fairlee Water Department, oral commun., 1993). Transmissivity calculated from aquifer-test data for well FLW-1 is approximately 13,200 ft^2/d, and average hydraulic conductivity is about 200 ft/d (table 3). Water from this well has high concentrations of iron and manganese. The town of Orford currently withdraws from a dug well (OSW-1) to supply water to a few homes. A campground on the banks of the Connecticut River withdraws from dug well (OSW-40) to supply water to campsites. The most productive location on the New Hampshire side of the aquifer may be at the northern end where Jacob's Brooks enters the Connecticut River on the floodplain.

Cottonstone Mountain Aquifer

The Cottonstone Mountain aquifer is west of Cottonstone Mountain and adjacent to the Connecticut River in western Orford (fig. 11, pl. 6). The Connecticut River Valley at this locality is narrow with very steep valley walls; glacial Lake Hitchcock was more than 600 ft deep here. A seismic-refraction line (Orford d-d', pl. 2, appendix C.30) 3,000 ft north of the Cottonstone Mountain aquifer shows the saturated thickness to be at least 600 ft, yet bedrock is exposed only a few hundred feet away at the valley wall across the river in Fairlee, Vt. About 800 ft southeast of seismic-refraction line Orford d-d', a homeowner attempted to have a well installed in bedrock but abandoned the project after the borehole was still in clay at 365 ft below land surface (OSA-1) (appendix A).

The Cottonstone Mountain aquifer consists of coarse-grained stratified drift with saturated thicknesses greater than 215 ft (pl. 6) and may be associated with a previous (pre-glacial) drainage channel of the Connecticut River. Two well drillers, attempting to install wells in bedrock, penetrated 215 ft of permeable sand and coarse gravel deposits at domestic well OSW-51 and 10 ft of sand and gravel deposits buried beneath 80 ft of fine-grained deposits at domestic

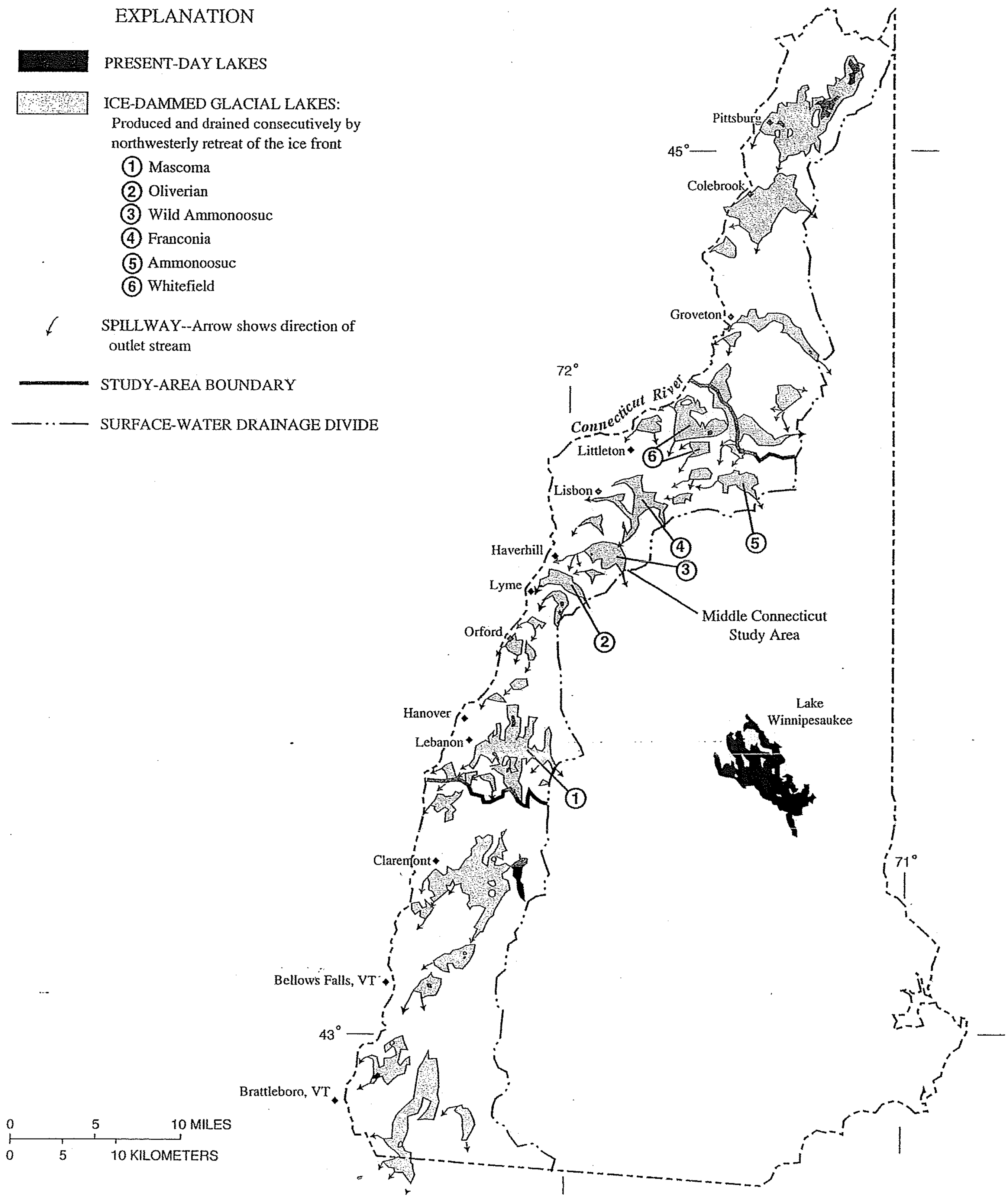
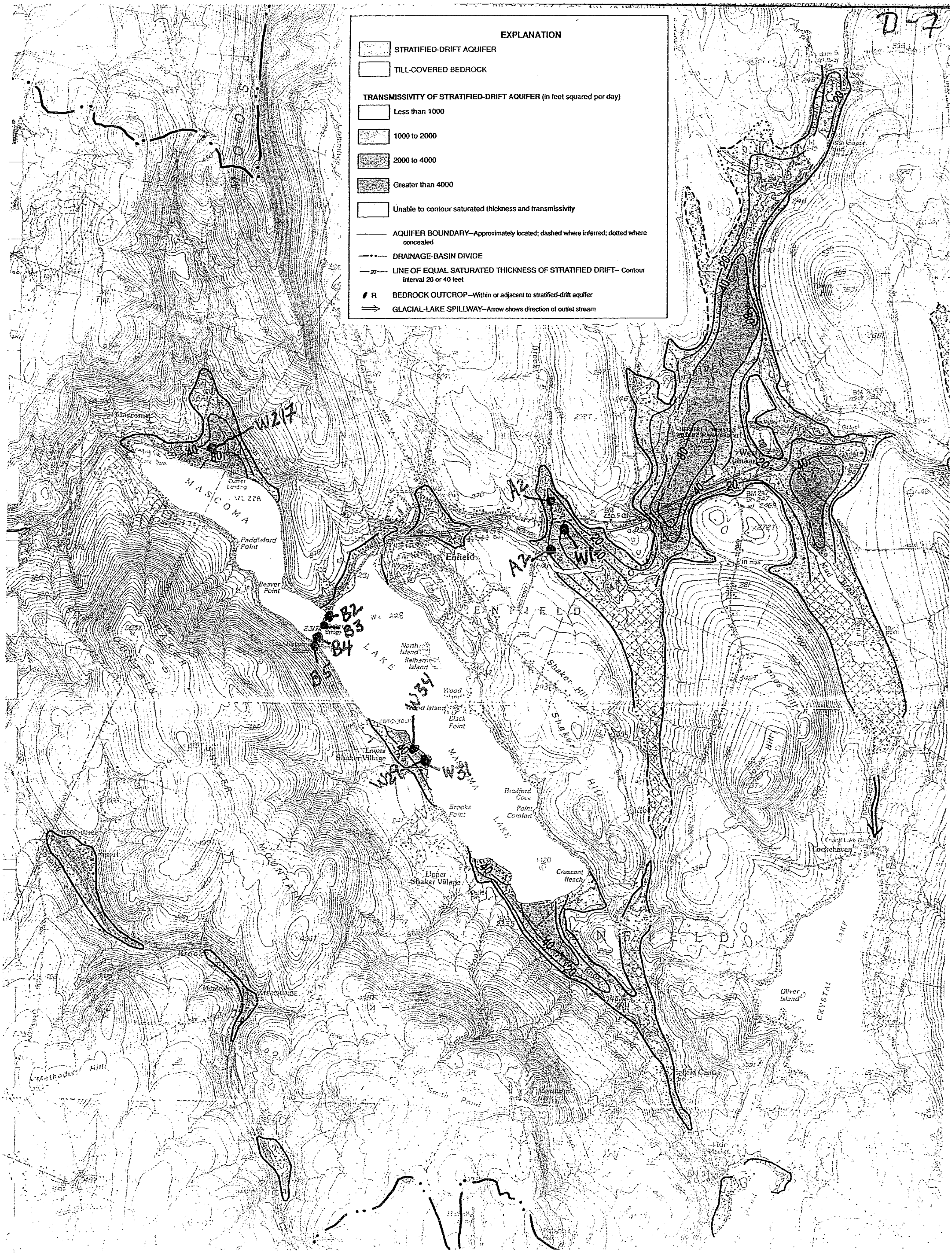


Figure 5. Approximate maximum extent of former glacial lakes in upland valleys in the Connecticut River Basin in New Hampshire. Only glacial lakes within the Middle Connecticut study area are identified; modified from Lougee, 1939)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EXPLANATION
STRATIFIED-DRIFT AQUIFER
TILL-COVERED BEDROCK
TRANSMISSIVITY OF STRATIFIED-DRIFT AQUIFER (in feet squared per day)
Less than 1000
1000 to 2000
2000 to 4000
Greater than 4000
Unable to contour saturated thickness and transmissivity
AQUIFER BOUNDARY—Approximately located; dashed where inferred; dotted where concealed
DRAINAGE-BASIN DIVIDE
LINE OF EQUAL SATURATED THICKNESS OF STRATIFIED DRIFT— Contour interval 20 or 40 feet
R BEDROCK OUTCROP—Within or adjacent to stratified-drift aquifer
GLACIAL-LAKE SPILLWAY—Arrow shows direction of outlet stream
W217
A2
W1
B2
B3
B4
B5
W34
W28
W31
MASCOMA
LAKE
Enfield
ENFIELD
Mascoma
Paddleford Point
Beaver Point
North Island
Wood Island
Black Point
Lower Shaker Village
Upper Shaker Village
Brooks Point
Bradford Cove
Point Comfort
Crescent Beach
Shaker Hill
Jones Hill
Lockehaven
Oliver Island
CRYSTAL LAKE
Methodist Hill
Smith Pond
West Canaan

Lithology	Depth (ft) From	Depth (ft) To
GRAFTON COUNTY—*Continued*		
Enfield		
Local site No. ENA 1		
Sand, fine, silty; some gravel	0	5
Clay, soft, blue	5	30
Hardpan, brown	30	--
Local site No. ENA 2		
Sand, fine, silty, brown	0	3
Clay, gray	3	14
Clay, soft, blue	14	33
Hardpan, brown	33	--
Local site No. ENA 3		
Sand, coarse and cobbles; many pebbles (250-500 mm)	0	12
Sand, medium to trace of very coarse sand, brown; mostly medium sand	12	18.5
Till	18.5	--
Local site No. ENB 1		
Sand, loamy	0	3
Gravel, sandy, slightly silty	3	9
Sand, silty	9	20
End of hole	20	--
Local site No. ENB 2		
Sand, coarse, and gravel, "firm"	0	31
Silt (fill)	31	33
Clay, blue, and sand	33	60
Sand and gravel "and mica"	60	61
Refusal	61	--
Local site No. ENB 3		
Sand, "loose," and gravel	0	2.3
Silt	2.3	9
Clay, blue, and sand	9	49
Sand and gravel	49	51
Refusal	51	--

Lithology	Depth (ft) From	Depth (ft) To
GRAFTON COUNTY—*Continued*		
Enfield—*Continued*		
Local site No. ENB 4		
Stones, loose (riprap)	0	8
Silt	8	13
Clay, blue and sand	13	54
Sand and gravel	54	56
Refusal	56	--
Local site No. ENB 5		
Riprap	0	3.1
Sand, fine "mud"	3.1	4.9
Sand, coarse, "loose, dirty," and gravel	4.9	26
Sand, silty, "soft"	26	30.3
Clay, blue, "soft," and sand	30.3	62
Clay and sand and gravel "mica"	62	64
End of hole	64	--
Local site No. ENW 1		
Clay, blue; some fine sand	0	17
Clay; some fine silt	17	20
Clay	20	40
Boulder (ice rafted?)	40	46
Clay	46	48
Interbedded lacustrine clays and coarse sand	48	77
Gravel, small (channel fill deposit?)	77	80
Lacustrine clays and coarse sand, interbedded (distal fluvial-deltaic deposit)	80	107
Hardpan, "stony," medium sand with compact silty clay matrix	107	116
Strongly foliated, biotite-rich schistose facies of Mascoma Group	116	
Local site No. ENW 29		
Sand, fine to medium, brown and gravel; some silt	0	14
Sand, fine to medium, brown, silty and gravel, fine to medium, broken, sharp	14	20
Sand, fine to coarse and gravel, broken, sharp, silty	20	35
Sand, fine and gravel, broken, sharp, silty	35	36
Sand, fine to coarse and gravel, broken and sharp; some silt	36	38
Sand, fine, silty; some gravel	38	39.8
End of hole	39.8	--
Local site No. ENW 31		
Sand, gray and gravel, broken, sharp	0	27
Clay, gray, and silt; few stones at bottom of well	27	48
End of hole	48	--
Local site No. ENW 34		
Sand, gray, silty	0	25
Gravel, broken, sharp	25	30
Clay, gray and silt	30	64
Gravel, broken, sharp	64	69
End of hole	69	--

Mascoma Lake Aquifers

Three aquifers are adjacent to the southern, southwestern, and northern shoreline of Mascoma Lake in Enfield and Lebanon (fig. 11, pl. 5). The first aquifer, underlying Knox River where it drains into the southern end of Mascoma Lake, covers 0.6 mi^2. The most potentially productive and thickest part of the aquifer, consisting of coarse-grained aquifer material overlying lake-bottom deposits, is in the center of the valley and close to the shoreline of the lake. The extent of these stratified-drift deposits beneath the lake is not known.

Seismic-refraction results (Enfield c-c', pl. 1, appendix C.18) indicate that the saturated zone is 50 to 68 ft thick (including till) in the center of the valley but that thickness of the aquifer generally is less than 40 ft. USGS observation well ENW-30, drilled at the western end of the seismic line, penetrated 34 ft of silty sand and gravel that changed at 50 ft to silt and clay lake-bottom sediments overlying till. Aquifer transmissivity in this area, estimated from the grain-size distribution of sediment samples collected during test drilling, is less than 2,000 ft^2/d. Most of this area is highly developed with summer and year-round homes with individual wells and septic systems. Most people living in the area obtain drinking water from the underlying bedrock aquifer.

The second aquifer is on the southwestern shoreline of Mascoma Lake near Lower Shaker Village. This relatively small aquifer area is currently (1993) pumped for public-water supply for Shaker Village (pl. 5). Public-supply well ENW-29 (pl. 1) is close to the shoreline of the lake and most likely induces water from the lake.

Seismic-reflection profiling along the shoreline of the lake near the second aquifer area indicated that saturated thickness exceeds 90 ft in places (fig. 8). None of the wells drilled during the siting of the new water-supply well reached refusal on till or bedrock. Public supply well ENW-29 penetrated 40 ft of silty sand and gravel; the proportion of silt in the sediments increased with depth. Transmissivity calculated from aquifer-test data for well ENW-29 ranged from 3,900 to 4,500 ft^2/d, and average hydraulic conductivity ranged from 115 to 132 ft/d (table 3) (Hydrogroup Inc., written commun., 1985).

Little data is available for the third aquifer, which underlies the northern end of Mascoma Lake near Mascoma Village (pl. 5). Seismic-reflection profiling along the shoreline of the lake near the aquifer indicated that the saturated zone is greater than 90 ft at the eastern end of the aquifer area (fig. 9). A domestic bedrock well (LHW-217) penetrated 30 ft of sand and gravel overlying 57 ft of clay deposits; bedrock was reached at 87 ft. Some of the other domestic wells in the area are shallow dug wells that tap the permeable sands and gravels overlying the lake-bottom sediments for drinking water. Aquifer transmissivity is probably less than 2,000 ft^2/d. If the aquifer is hydraulically connected to the lake, water-supply wells drilled near the lake could potentially increase yields by inducing recharge to the aquifer; however, dense residential land use in the area limits potential of the aquifer for production of drinking water because of possible water-quality concerns.

Goose Pond Aquifer

Goose Pond aquifer, formed in glacial Lake Mascoma, is in West Canaan (fig. 11, pl. 5) and underlies a 3.1 mi^2 area. The most potentially productive part of the aquifer area consists of glaciofluvial aquifer material confined beneath clay and peat deposits that formed in glacial Lake Mascoma.

In one area of the Goose Pond aquifer, the saturated thickness of the confined, coarse-grained aquifer material is 27 ft (USGS observation well CCW-7; pl. 1, appendix B). USGS test boring CCA-3 penetrated multiple layers of coarse-grained aquifer material interlayered with lacustrine silt; glacial till was reached at 65 ft. Seismic-refraction data in the area (Canaan lines a-a', b-b', c-c', d-d'; pl. 1, appendixes. C.4-C.5) indicate that the saturated zone in the Goose Pond Brook Valley ranges from 47 to 84 ft (the shallowest part is toward the eastern end of the valley). At the southern end of the aquifer near Herbert Webster Wildlife Management Area and New Hampshire Route 4, aquifer thickness