Glacial Tectonism at Hither Hills, Long Island, Revealed by Ground-Penetrating Radar Investigations

Kurt Goetz
MS Thesis 2005

Department of Geosciences
Stony Brook University
Stony Brook, NY 11794-2100
Introduction

An actively advancing glacial ice sheet can have an enormous impact upon the terrain it comes in contact with, indiscriminately depositing, eroding, and deforming sediments it comes upon. It can transfer large amounts of stress through loading, compression, and shearing via a variety of processes referred to by the sweeping term *glacial tectonism*. Active glacial margins worldwide, from both ancient and modern glaciations, have been responsible for the formation of unique landforms. This report focuses upon features in far eastern Long Island, New York, which resemble in outward appearance some of these landforms, and possible hypotheses for their formation as a part of the genesis of Long Island.

This study is of profound interest for the region. The area is solely reliant on groundwater for the production of natural potable water. An understanding of the formation of the area might therefore provide additional information regarding the distribution of aquifers without the need for costly, time-consuming, and extremely localized well measurements. In addition, the study could be used to indicate the history of the glacier which formed the area, aiding glaciological and climatological studies. Finally, it could present a model which can be applied to modern and past glaciers in explaining features unique to glacial processes.
The seemingly inconsequential loading of ice creates a unique and remarkable deformational regime at and near a glacial margin. While the magnitude of the stresses transmitted to the underlying crust may be relatively insubstantial compared to those required for continental-scale orogenesis, the rate of loading and unloading takes place on such a short time scale (several orders of magnitude faster than orogenesis) that deformational features are quite common near the glacial snout, in some cases to depths of bedrock and beyond. In general, the processes are similar in large-scale continental tectonism and glacial tectonism despite the difference in scale (Sharp et al., 1984). Indeed, Thorson (1996, p. 1187) notes that mechanically, “each glacial advance can be considered as a rapidly propagating, gravity-driven detachment fault.”

*Glaciotectonic Landforms*

There is a variety of ways in which positive topographic landforms can be created by imposition of stresses (Fig. 1). Rather than the typical meltout and recessional moraine modes of pure deposition, these landforms (referred to as glaciotectonic) involve a kind of small-scale tectonism. Any horizontal stress component sufficient to overcome the soil and ice overburden and the material’s yield strength will cause a vertical displacement of material beneath, beyond, and within the ice. As glaciers impose a wide range of stresses (most significantly pushing, vertical loading, and subglacial shearing) upon the glacial environment, a similarly wide range of deformational
Figure 1: Selected modes of glaciotectonic construction of margin-parallel ridges by active glaciers. **A:** Squeezing-out of moraine from beneath the glacial margin due to overpressuring of saturated sediments. Squeezing upwards into basal crevasses is also possible. From *Hart*, 1998. **B:** Freezing-on and rafting of intact portions of glacial substrate to be deposited *en masse* upon stagnation and melting of the glacier. From *Weertman*, 1961. **C:** Push moraines formed by the pushing of unconsolidated sediments into folded mounds. From *Hart and Watts*, 1997. **D:** Modern push moraine (dark, dirty material) at active margin of Hofdabrekkujökull, Iceland. Note surveyors for scale. From *Krüger*, 1985. **E:** Thrust ridges formed at the front of individual imbricate critical wedge-style thrust sheets. After *Hambrey*, 1994.
features can be formed at or near the margin. The observations of glaciotectonic landforms from field studies on modern glaciers (although relatively sparse, due to the scarcity of modern advancing glaciers, as pointed out by Krüger (1985)), combined with sedimentological and structural analysis of paleoglacial landforms, have increased our understanding of the processes involved in creating them.

Subglacial shearing and loading (pressing), especially involving saturated, unconsolidated sediments beneath temperate glaciers (those with soles at the pressure-melting point), can cause fluidization and pressure-driven flow of the basal material upwards, creating small diapiric ridges wherever an outlet presents itself. Sediments may be extruded in front of the glacial margin (Fig. 1A: e.g., Price, 1970; Sharp, 1984) or upward into radial or margin-parallel basal crevasses (e.g., Andrews, 1964; Hart and Smith, 1997), creating ridges which are typically small, horizontally discontinuous, steeper on their distal sides, and characterized by a strongly preferred till fabric perpendicular to the trend of the ridge (which may not represent the former direction of ice flow) (Andrews, op. cit.). Such a formation process has been postulated for the formation of De Geer moraines in Scandinavia (by a Younger Dryas tidewater margin) (Lundqvist, 1988) and similar features in Baffin Island (Andrews, op. cit.) and Iceland (Price, op. cit.). These features will be referred to as “squeeze moraines.”

For a cold-based glacier (or more appropriately, one whose base oscillates on either side of the pressure-melting point) encroaching upon a similar environment, layers of basal material may “freeze onto” the base of the glacier (Fig. 1B: sensu glacio-erosional rafts of Ruszczyńska-Szenajch, 1987). Meltwater produced during the melting-base phase of such a glacier penetrates the substrate, provided it is permeable enough to allow such penetration (subglacial saturated
lodgment till and/or overridden glaciofluvial deposits readily fit this description). During the ensuing cooling period, this water refreezes within the sediment, forming interstitial ice which adjoins to the main glacial body, and the material becomes incorporated intact into the glacier, in some cases in slabs up to the better part of a meter thick (Krüger, 1993). Transport then continues englacially, although slabs in this part of the glacier may be dissected and/or thrust farther vertically into the glacier when internal thrust sheets develop in the ice (Fig. 1B; Weertman, 1961) or basally (e. g., Waller et al., 2000). The entrained material can then be deposited into mounds upon stagnation and melting of the glacier, at times displaying bedding that is almost undeformed from the pre-entrainment state. The external form of these mounds is often controlled by the underlying topography (Krüger, 1995). Krüger (1993, 1995, 1996) has conducted an in-depth study of annual moraines constructed in this manner in front of the Icelandic glacier Myrdalsjökull, and Weertman (op. cit.) uses this process to explain similar features in both Arctic and Antarctic regions.

The marginal and near-proglacial environment is the locus of significant stress gradients and the juxtaposition of contrasting stress fields (Boulton and Caban, 1995), with subglacial extension due to shearing abruptly decreasing as the margin is approached, replaced by proglacial compression (e. g., Croot, 1987 Fig. 7; Hart et al., 1990 Fig. 2). These stresses are superimposed upon the decreasing vertical load provided by the downglacier-thinning snout. In addition, recent glaciofluvial, supraglacial meltout, and relatively unconsolidated till deposits near the margin create a ready canvas upon which the glacier can construct its tectonic mural. The combination of these characteristics makes the glacial margin an ideal location for the development of glaciotectonic landforms.
Loose proglacial materials may be pushed by an advancing glacier (whether overall or seasonally in an overall retreat phase) into small mounds in a manner controlled by the critical wedge principle of *Davis et al.* (1983) (Fig. 1C): these features will therefore be steeper on the proximal side. Modern features caused by this process have been recorded in Iceland by *Krüger* (1985; Fig. 1D), while *Ham and Attig* (2001) documented seasonally-developed 1-4 m high ridges from the Wisconsinan Ice Sheet in Pleistocene Wisconsin. These features will be denoted herein as “push moraines.”

More consolidated sediments may be compressively folded into ridges, as described in Chile by *Schlüchter et al.* (1999), or thrust as coherent blocks along a basal décollement (Fig. 1E). Both of these forms are typically arcuate and follow the trend of the former ice margin. Once the gravitational stresses created by the uplift reach a critical threshold (determined by a combination of glacial, hydrological and sedimentological criteria), it becomes more efficient to transmit the compressive stress farther into the glacial foreland horizontally than to continue uplift, and the décollement propagates forward along a new, shallower plane (*e. g.*, *Elliott*, 1976; *Boulton et al.*, 1999). Multiple, imbricate thrust sheets of basal material can be generated in this manner if the glacial advance continues “bulldozer-style” into the glacial foreland (*sensu Davis et al.*, 1983), with a ridge being generated at the nose of each sheet (Fig. 1E; *Croot*, 1987, Fig. 5). These two mechanisms are almost identical to those involved in the generation of fold-and-thrust belts in continental orogens (*e. g.*, *Dahlstrom*, 1970; *Elliott*, 1976; *Boyer and Elliott*, 1982). The landform produced specifically by thrusting is herein referred to as a “thrust moraine.” Thrust moraines have been observed at a variety of scales in front of modern glaciers on Svalbard Island (*Hambrey and Huddart*, 1995; *Huddart and Hambrey*, 1996; *Boulton et al.*, 1999) and Iceland.
(Krüger, 1985; Croot, 1987), and as Wisconsin-age large-scale features in western Canada (200’ high, 600’ wide ridges: Kupsch, 1962).

A combination of processes may also be at work in a glacier, both spatially and temporally, producing a variety of deformational structures (Boulton et al., 1999). As an example, Eybergen (1987) described the initial stacking of thrust sheets at and just beyond the margin of the modern Swiss Turtmannglacier as it advanced. During further advance, this material was overridden; a portion was extruded outwards above the thrust sheets, and another portion was frozen onto the base and deposited atop the section. Intense shearing subsequently deformed the entire segment, while new folds and thrusts formed at the new marginal position (see his Fig. 10).

Both thrust and push moraines obviously require glacial advance for their formation. In this regard, they can be used (to a certain extent) as paleoclimatic indicators, in that they denote an overall advance period (e. g., Boulton, 1986) and perhaps the start of an abrupt warming trend (Lowell et al., 1999). They may be formed as the culmination of a single large-scale advance (e. g., Hart, 1990) or seasonally, with yearly readvances coinciding with the winter season, when accumulation outpaces the marginal ablation (e. g., Worsley, 1974; Sharp, 1984). Sharp (1984) was able to draw a correlation between the spacing of annual moraines at a modern glacial margin in Iceland and the recorded rates of its retreat (Fig. 2). This suggests that glaciotectonic ridge forms have the potential to provide a certain amount of paleoclimatic information (see Lowell et al., 1999).

Potential Role of Meltwater and Pore Fluids

A substantial amount of meltwater is produced near the temperate glacial margin, even in the case where the glacier is undergoing a net advance. Coupled with the hydrologically
Figure 2: Observed rate of retreat of Skálafellsjökull, Iceland (●), compared with spacing between corresponding recessional moraines (□). From Sharp, 1984.
heterogeneous nature of the glacial foreland, the differential flow of meltwater is capable of abetting the processes of glaciotectonic deformation in a variety of ways.

The most straightforward manner in which meltwater can assist in glaciotectonism is by generating unconsolidated sediments to be deformed through sub- and englacial erosion and deposition at or near the margin. Subglacial meltwater channels, sheetflow, and flood events (e.g., Alley, 1988; Shoemaker, 1992b; Brennand and Shaw, 1994; Beaney, 2002) are particularly effective at eroding the glacial substrate and transporting the sediment to the margin, driven by substantial pressure gradients. Coarse outwash is typically deposited near the glacial margin, while finer sediments are carried farther as the glacial streams lose competence. If a large topographic feature (such as another glacial lobe, a chain of mountains, or an end moraine) is present some distance from the glacial margin, a proglacial lake can develop between the ice margin and the obstacle through damming, behaving as a sediment trap in which easily-deformed glaciodeltaic and glaciolacustrine sequences are created. In addition, supersaturated till at and just beyond the margin can become fluidized and mobilized under its own weight (Owen, 1987), forming unconsolidated deposits variously referred to as “flow till” (Morawski, 1985) or “mud flows” (Boulton and Deynoux, 1981).

Meltwater also has a significant influence on glaciotectonic mechanisms by lowering the effective stress due to the combined overburden of the sediment and the glacial ice, thereby lowering the strength of the sediment. The particles of material are confined by the hydrostatic pressure $P$ due to the combined stresses of the soil and ice overburden, which is sufficient to preclude deformation under dry conditions. However, the fluid pressure $P_w$ caused by intrusion of pore fluids acts to
support a portion of the overburden, lowering the hydrostatic pressure to an equivalent effective pressure $P_e$:

$$P_e = P - P_w$$  \hspace{1cm} (1).

Because this effective confining pressure $P_e$ can be considerably smaller than the hydrostatic stress $P$, there is a significant reduction in the shear strength $\tau$ of the substrate material, which is related to the confining pressure in a simple Coulomb relationship:

$$\tau = C + P_e \times \tan(\phi)$$  \hspace{1cm} (2),

where $C$ is the cohesion of the material, typically taken as 0 in unconsolidated granular material (but considerably higher in permafrost material, as will be noted), and $\phi$ is the angle of internal friction (e.g., Hubbert and Rubey, 1959; Owen, 1987). Combining (1) and (2), it becomes clear that supersaturated substrate material will be weakened due to elevated values of the fluid pressure $P_w$, allowing deformation. As $P_w \to P$, $P_e \to 0$ and the material will essentially “float” and be entirely supported by pore fluids rather than by intergranular contact. In this case, deformation can be initiated at remarkably low driving stresses (e.g., van der Wateren, 1985): in a larger-scale environment, analysis by Hubbert and Rubey (1959) has shown that increased pore water pressure is capable of allowing massive blocks of earth to be overthrust tens of kilometers from their source by gravitational sliding alone. The significant amount of meltwater beneath much smaller glaciers surely can have a dramatic effect on deformation processes.

The spatial variation of pore water pressure depends on a number of criteria. The presence of an impermeable barrier at a relatively shallow depth, such as a clay layer (Banham, 1975; van der Wateren, 1985) or condensed lower till horizon (Boulton et al., 1974; Boulton and Hindmarsh, 1987), will act as a vertical confining layer and create a substantial buildup of water.
pressure and a probable décollement above this layer. As will be described below, the presence of large amounts of permafrost or interstitial ice has a similar effect by decreasing the potential for groundwater escape. The distribution of unconsolidated sediments also determines the direction of groundwater flow. The rate of glacial advance is also significant: if the glacier can outpace the subglacial groundwater flow, there will be a substantial increase in glacial loading, increasing $P$ (and therefore $P_e$ by (1)), but also preventing the vertical escape of the water and lowering $P_e$ more substantially (van der Wateren, 1985; Aber, 1988).

Potential Role of Permafrost

More controversial is the role of permafrost and/or interstitial ice in various glaciotectonic processes. The key property of permafrost of interest to the present study is its enhancement of sediment cohesion $C$ in (2), which can apparently be detrimental or beneficial to glacial tectonism, depending on the exact mechanism involved.

The presence of permafrost significantly increases the strength of unsaturated sediments, so much so that the glacier ice will preferentially deform before the sediments in most cases (Nickling and Bennett, 1984; Etzelmüller et al., 1996). This relationship holds only for essentially instantaneous loads: the viscosity and shear strength of frozen soils over longer-term (years) application of load has been shown by Tsytovich (1975, p. 37, 117-123) to decrease markedly, easing deformation of sediments. However, at the relatively short time-scale upon which glacial tectonism acts, the deformation of ice (both in soils and within the glacier itself) before sediment creates the unusual effect of reversing the usual grain size-deformational competence relationship: interstitial ice between granular materials such as sand and gravel can act as a potential thrust
surface and preferentially deform, while the micro-scale structure of clays is such that they behave in a mechanically competent manner when frozen (Croot, 1988; G. Hanson, personal communication, 2004). This affects the rheology of sediments and hence the style of glaciotectonic deformation in cases of a glacier advancing into a permafrozen foreland.

Such an increase of sediment strength would seem to make glacial tectonism more difficult under typical circumstances, requiring more stress to overcome the increased cohesion of the sediments (Thomas, 1984). However, this stronger material can also be more efficient at transmitting horizontal compressive stresses into the glacial foreland, in some cases making proglacial deformation more likely (Boulton et al., 1999). In certain cases, it can allow the overthrusting of coherent sediment blocks over the scale of many kilometers away from their sources (Fig. 3: Meyers et al., 1998).

The aforementioned effects of porewater pressure are also affected by permafrost in an ambiguous manner. The results of flowline modeling analysis (Boulton et al., 1995; Boulton and Caban, 1995) suggest that permafrost can act as an impermeable proglacial surface, increasing the area of elevated subsurface pore water pressure (and hence decreased sediment strength) beyond the margin (Fig. 4). This has been used to explain deformation in paleoglacial features in the Yukon (Mackay, 1960) and the Netherlands-Germany region (Boulton et al., 1995). A counter-argument is offered, however, by Thomas (1984) and Schlüchter et al. (1999), who suggest that any pore fluids in a permafrost region would likely be frozen into the sediments, preventing a decrease of the effective pressure. Frozen sediments would also prevent the formation of soft-sediment deformation features, such as squeeze moraines (Andrews, 1964; Aber, 1988).
Figure 3: Glacially transported sediment blocks, Ranco, Long Island. Units 3 and 4 were transported en masse while permafrozen: thrust sheets developed in sands and gravels (see text for explanation). Site is on Ronkonkoma moraine, illustrating potential for permafrost in study area of present report. Photo and interpretation after Meyers et al., 1998.
Figure 4: Effects of permafrost on Pw. Difference in pressure-driven groundwater flow near the glacial margin due to the presence (a) and absence (b) of permafrost layer. From Boulton and Caban, 1995.
However, Tsytovich (1975, p. 25) notes the potential for the presence of at least some liquid pore water at temperatures as low as -70° C, offering the potential for fluid-related processes.

The increased cohesion of sediments also eliminates the possibility of folding sediments in a ductile manner, but increases the likelihood of brittle thrusting, with the boundary between the permafrost and unfrozen sediments below potentially behaving as a décollement (Aber, 1982). In some cases, the thickness of glacially-thrust sheets of sediments has been used as a proxy to determine the depth of permafrost at the time of deformation (e. g., Ingólffsson, 1994). Very thick permafrozen layers can also act as an abutment against which thrust sheets can be stacked to form large ridges (e. g., Lehmann, 1993).

Krüger’s (1993, 1995, 1996) basal freeze-on mechanism requires a lack of permafrost, as meltwater must percolate seasonally into the substrate. Alternatively, other studies require only partial permafrost to develop glaciotectonic features (Thomas, 1984; Etzelmüller et al., 1996), the contrast in cohesion creating a variety of deformational styles in similar sediments. Finally, many types of glaciotectonic landforms are able to form in the presence or absence of permafrost (e. g., Humlum, 1985; Ingólffsson, 1994; van der Wateren, 1995). Due to these ambiguous roles of permafrost in the various glaciotectonic processes, very little information on the importance of permafrost can be drawn from the present study. In addition, while there are indicators that permafrost was present south of the Wisconsinan Ice Sheet’s southern margin (e. g., New Jersey (Wolfe, 1953); Ranco Quarry, along the Ronkonkoma Moraine in Manorville, Long Island (Meyers et al., 1998: see Fig. 3), no indicators have been found in my study area.
Glaciotectonic processes have become increasingly invoked in the literature to describe orderly features in known glaciated areas, particularly near former marginal positions. The present study focuses on a feature in Long Island, New York (Fig. 5) which is located on a known marginal position of the Wisconsinan ice sheet. The feature in question is hypothesized to represent a glaciotectonic landform complex. Evidence in support of this hypothesis is based upon its topography, its location relative to a moraine, and previous stratigraphic and geophysical field studies.

Glacial Tectonism on Long Island

For more than a century, glaciotectonic features have been noted throughout Long Island. Merrill (1886) described folding and faulting in the Harbor Hill and Ronkonkoma moraines, the large ridges forming the “backbone” of the north and south fork of the island, respectively, which are thought to have been formed by the margin of the Wisconsinan-age Laurentide Ice Sheet during its presence in the New York/New England region. He ascribed the folding to glacial pushing during advance of this sheet. Similar deformation has been recorded on Block Island, Martha’s Vineyard and Nantucket Island, which have all been postulated to represent eastward extensions of the Ronkonkoma Moraine (e.g., Kaye, 1964a, b; Sirkin, 1976, 1982; Oldale, 1982; Stone and Borns, 1986). Fuller (1914) noted a number of glaciotectonic features,
Figure 5: Study area.  A: Map of Long Island showing location of Hither Hills region.  B: Aerial photo of Hither Woods area.  NE-SW trend of hills is visible.  Also note dendritic valley network and north-trending Power Line Cut (PLC) in center of highlighted area.  C: Topographic map of Hither Hills State Park area, displaying NE-SW trend of hills and PLC in red (see Fig. 27 for profile.)
particularly in the form of folding, throughout Long Island in both the Ronkonkoma and Harbor Hill moraines due to what he interprets as a mid-Pleistocene (Illinoian) ice sheet, each moraine supplemented by further deformation by a later (Wisconsinan) ice sheet (p. 201-211). The prominent coastal bluff exposures on Long Island’s north shore show extensive evidence of glacial folding and thrusting (Fuller, 1914; Mills and Wells, 1974; Meyers et al., 1998; Selvaggio and Richard, 1999), and subsurface fold-and-thrust deformation in the Stony Brook lobe of the Harbor Hill Moraine has been observed geophysically (Tingue et al., 2004). In the Ronkonkoma Moraine, glaciotectonic thrusting has also been recorded near Manorville by Meyers et al. (Fig. 3: op. cit.), and evidence of deformation continues sporadically eastward throughout the south fork of the island (e.g., Fuller, 1914; Newman et al., 1968; Nieter et al., 1975; Black et al., 1998).

**Hither Hills State Park**

Hither Hills State Park (Fig. 5) is part of an abnormally wide section of the isthmus between Napeague State Park and Montauk Peninsula on the eastern end of the south fork of Long Island. The region consists of Hither Woods Preserve and the Hither Hills County Park Preserve to the east, and Hither Hills State Park and the “Walking Dune Field” to the west. The area is located within the town of East Hampton, 12 km west of Montauk Point, and forms a part of the Ronkonkoma Moraine. The present study was largely conducted on and near the eastern boundary of the state park. The local area as a whole will be referred to as the Hither Woods (rather than Hither Hills, to avoid confusion with the state park proper or the physical landforms which are the subject of the study).
Within the Hither Woods is a series of hills trending roughly ENE-WSW with a relatively uniform spacing (Fig. 5B, C) over an area extending 5 km east-west by 2 km north-south, referred to here as the Hither Hills. The hills typically are linear features traceable for 250-750 m along strike, with an across-strike wavelength generally around 100 m. Their height from crest to trough ranges from 3-13 m, and they exhibit a large-scale slope with the highest altitude to the south. The orientations of the hills seem to fall into small clustered groups, hills in each group generally showing a consistent (within about 5°) trend (Fig. 6; Bernard, 1998).

Located about 2/3 of the way north of the study site is a network of valleys which approximates a dendritic pattern (Fig. 5B, C). This unusual feature includes the lowest topography in the Hither Woods and invites comparison to a fluvial erosional process. This area is the study site of an independent geophysical survey by Mitchell Cangelosi, and some of his data and conclusions, where relevant to this study, will be included herein.

The origin of Hither Hills has been explored occasionally by previous workers. Fuller (1914, p. 47-48) interpreted the linear valleys between ridges as eroded runoff channels that collected water flowing off of the moraine. Sirkin (1982; 1995, p. 46) described the arcuate ridge system as the result of repeated, passive recessions and stagnations of an ice sheet margin during an overall retreat phase. More recently, glaciotectonic deformational hypotheses have been put forward by Bernard (1998; Bernard et al., 1998) and Klein, Davis and colleagues (Klein and Davis, 1999; Klein et al., 2000), and Fuller (op. cit., p. 48) left open the possibility that the runoff channels in his interpretation may have followed the path of folded valleys.

A network of hiking trails running throughout the Hither Woods allows a degree of accessibility to the region (Fig. 5C). In addition, the Long Island Lighting Company (now the Long Island Power Authority) installed a series of high-voltage lines with regularly spaced (250
Figure 6: Orientation of Hither Hills. Hills fall into small clusters with similar (within 5º) orientations. Each cluster might represent an individual marginal position of Wisconsinan Glacier or partial overriding episodes creating a “palimpsest” terrain. From Bernard, 1998.
feet) utility poles stretching along a roughly north-trending line which cuts directly across the full width of the park on its eastern boundary, with an accompanying vehicle maintenance trail (Fig. 5B, C). The trail allows relatively easy access to the site. It branches directly off of Montauk Highway and provides a favorable location for a radar survey, as it opens a remarkably linear path directly across the south fork. This feature is hereafter referred to as the Power Line Cut (abbreviated “PLC”). The first hill along this trail was the central location for the seismic surveys of Bernard (Bernard, 1998; Bernard et al., 1998), allowing a direct comparison with the results from that survey, whose findings will be related shortly.

The utility poles which once carried the high voltage lines along the PLC have since been cut down, and a subsurface pipe placed in their stead. The 29 regularly-spaced, meter-high stumps remaining from the poles are useful in calibration of the survey lines and are occasionally used as markers to begin individual lines. However, the subsurface pipe creates a number of complications, to be outlined in the Methods section.

There are a number of characteristics of the Hither Hills which suggest that their origin is glaciotectonic. The regular curvilinear (“washboard”) arrangement of the hills (Fig. 5B, 6) suggests that they might represent margin-parallel topographic features related to the retreat and/or readvance of the Wisconsinan glacier which formed the Ronkonkoma Moraine. Construction could have taken place as purely depositional features (i.e., recessional moraines) or as one of the myriad glaciotectonic varieties described in the Previous Works section; however, the relatively extreme curvature (in cross-section) and topography of the hills seems to favor a glaciotectonic origin, since less-compact recessional moraines, typically composed of chaotic debris assemblages, would tend to collapse into more stable configurations shortly after deposition. The substantial, rolling topography, particularly in its steepness and along-strike regularity, is unique on Long Island.
Recently, substantial visible geological evidence has been discovered to support the glaciotectonic theory. A large-scale exposure was revealed during the excavation of a landfill in the southeastern portion of the Hither Woods several decades ago, revealing the presence of a large overturned antiform and a potential thrust sheet (S. Engelbright, personal communication, 2001). Also, a number of exposures in the coastal regions of the area reveal further evidence of glaciotectonic processes. A large-scale series of folded diamicts was recorded in exposures in a small bluff along the northern shore of the park (Fig. 7A) by Klein and Davis (1999). Similar coastal exposures have since been revealed and documented during the course of my field surveys (Fig. 7B, C). The deformation observed was particularly striking in one particular exposure which the erosional coastline has cut at an orientation oblique to the strike of the hills (most of the exposures are parallel to this strike). The aforementioned authors noted the extreme variability of the stratigraphy of these and similar exposures (Fig. 7D): the glacier responsible for the deposition and/or deformation of these sediments was likely deforming different packages of sediment in each hill. The folded and organized character of these outcrops is clearly inconsistent with a simple depositional (recessional moraine) model, as the scale and intensity of folding cannot be explained by slumping alone, nor is there a simple stratigraphic explanation for the sediment variability.

Finally, the internal structure of the hills is strongly supportive of a glaciotectonic origin. The exposures noted in the preceding paragraphs are serendipitous occasions: outcrops on Long Island are typically found only in temporary quarries and storm- and wave-cut bluffs, and the easily-eroded sediments do not preserve revealed structures long enough for repeated visits (Fuller, 1914; Black et al., 1998; Bernard et al., 1998; Meyers, 1998). Therefore, various
Figure 7: Stratigraphic evidence of glacial tectonism at coastal exposures on northern shore of Hither Woods.  

A: 1998 photomosaic of coastal exposure revealing large-scale glaciotectonic folds in sand.  

B, C: 2005 photographs of same exposure.  Further coastal erosion has revealed more internal detail of deformation.  

D: Stratigraphic columns of exposures at Rocky Point area, northeast coast of Hither Woods, by Klein and Davis (1999) showing variability of sediment composition in a small area: Glacier deposited and deformed a different package of sediments with each readvance.
-Photo by Elliot Klein
subsurface geophysical techniques in the Hither Woods area have been undertaken by our research group to establish the subsurface structure *in situ* and non-invasively. The seismic surveys of Bernard (1998; Bernard *et al.*, 1998) in the southern portion of the park revealed a folded surface interpreted to be glacially thrust, varying from 4-13 m below the surface (Fig. 8). Results from the Bernard survey were unfortunately hampered by the uncompacted, sandy nature of the moraine deposits, as unconsolidated materials tend to attenuate seismic waves because of energy loss through intergranular contact and compaction (Bernard, 1998). Preliminary radar studies near the sites of the coastal exposures (Klein *et al.*, 2000) and the seismic surveys (Winslow *et al.*, 2000) imaged similar folded layers at a comparable depth. The subsurface deformation appeared to be of similar scale to the exposed folds on the coastal bluffs (compare with Fig. 7A, B), indicating perhaps a similar or single origin for the entire chain of hills.

The present study is intended to draw upon the experience of previous studies in the Hither Woods, and to correlate the indisputably glaciotectonic deformation observed in the coastal exposures with the geophysically imaged reflectors beneath the rolling hills. Further, it is my intent to obtain data of higher quality and over a wider area than the previous studies, to determine the style and amount of deformation in a large portion of the area. Implications from these observations are intended to aid in reconstructing the glacial processes responsible for the formation of the Hither Woods and, if appropriate, the entire south fork of Long Island.
Figure 8: Previous seismic evidence for glacial folding. Common-depth-point stacked seismic reflection section along southern hill in Hither Hills State Park from Bernard (1998), showing reflective layer about 4-13 m below the surface. Layer interpreted as glacially deformed stratigraphic layer.
Ground-Penetrating Radar

Ground-Penetrating Radar (GPR) is a geophysical exploration tool which is commonly used for the lateral and vertical mapping of subsurface features according the manner in which they reflect electromagnetic (EM) waves. It is non-invasive, can be applied to a variety of geologic environments, and collects data fairly rapidly, making it a versatile and incrementally inexpensive geologic tool. It can also resolve features on a decimeter scale or can penetrate to depths of many tens of meters. However, it should be stressed at the onset that the data collected often do not lend themselves to a unique interpretation, and typically must be augmented with data acquired by other proven geologic techniques (e.g. boreholes, outcrops, lysimeters, other geophysical methods: see Stoffregen et al., 2002) to obtain a meaningful and robust interpretation.

GPR has been applied in groundwater and hydrological exploration (e.g., Beres and Haeni, 1991; Tsolias et al., 2001; Hubbard et al., 2002; Stoffregen et al., 2002) and contaminant plume monitoring (e.g., Brewster & Annan, 1994; Sandberg et al., 2002). It has been applied elsewhere in previous studies to study glaciotectonic features, both directly in modern, active glacial environments (e.g., Lønne and Lauritsen, 1996) and in paleoglacial deposits (e.g., van Overmeeren, 1998; Busby and Merritt, 1999; Overgaard and Jakobsen, 2001; Jakobsen and Overgaard, 2002; Tingue et al., 2004), illustrating its flexibility and applicability.
Figure 9: Basic principles of Ground-Penetrating Radar. The transmitting antenna sends out a wavelet of hemispheric electromagnetic energy (here represented by a single raypath), a portion of which is reflected by subsurface media of differing relative permittivity and recorded by the receiving antenna. As the antennae are moved along a traverse in tandem (in Common Offset configuration), a 2-dimensional subsurface map of two-way wave travel time vs. horizontal survey distance (radargram: Fig. 10) is constructed. Modified from van der Kruk et al., 1999 and Hubbard et al., 2002.
to multiple geological environments. However, the present study, to the best of my knowledge, is
the largest-scale use of GPR in a glaciotectonic study to date.

The GPR system consists of a pair of antennae (Fig. 9) which operate on electromagnetic
(EM) frequencies in the radar spectrum. A transmitting antenna sends out a short hemispheric
wavelet of energy focused around a nominal peak frequency, f. Frequencies used in GPR surveys
typically range from 25-1000 MHz. The outgoing wave is transmitted through the subsurface, is
partially reflected off of subsurface media of differing EM properties, and returns to the surface to
be recorded by a receiving antenna which is triggered a short time after the transmitter. The
receiving antenna records the amount of travel time required for reflected portions of the signal to
return. If the EM wave velocity of the subsurface material can be determined (or approximated
adequately), this two-way travel time can then be used to determine the depth of the object which
caus[146x643]ed the reflection.

The frequency of the particular radar antennae determines the resolution and depth of
penetration the survey is able to achieve. The frequency f of the signal is of course inversely
proportional to its period, p. The characteristic wavelength $\lambda$ of the signal is determined not only
by the particular antennae used, but also by the subsurface medium: specifically, the wavelength is
determined by the product of the subsurface velocity $V$ and wave period, so that:

$$\lambda = \frac{V}{f}$$  \hspace{1cm} (3).

Accordingly, high-frequency surveys are characterized by relatively short wavelengths. This means
that the waves are more sensitive to smaller objects, giving them better resolution (resolution is
typically approximated as $\frac{1}{4}\lambda$). However, higher-frequency waves are more vulnerable to various
types of signal attenuation, and therefore are not able to penetrate to great depths. Conversely,
low-frequency surveys are able to penetrate to much greater depths, but their large wavelength (equation 3) leads to a loss of resolution. The scale of the target(s) of each survey therefore must be thoroughly understood in order to optimize the survey.

The key factor controlling the propagation velocity $V$ of the medium through which the EM waves are traveling is the dimensionless *relative permittivity* (also known as the dielectric constant) $\varepsilon_r$ of the medium, a parameter which determines the velocity of the medium relative to that of EM radiation in a vacuum, $c$:

$$V = c \times (\varepsilon_r)^{-1/2}$$

(4).

This relation holds only for nonmagnetic, nonconductive media, but it is a good approximation for the till material of the Hither Woods, as it has negligible conductivity and very minor if any magnetic sand content. Air has a relative permittivity of 1, allowing waves to propagate at $c$, while sand has values which fall between about 4 and 25 depending on moisture content, with higher values representing wetter material (*Davis and Annan*, 1989). This difference becomes important in interpreting the radar results, as velocity is frequently inhomogeneous due to small- to large-scale spatial differences in mineral, air, and water content of the medium. *As Bernard* (1998) points out, judging by the scale of the hills, any detachment structure is probably shallow enough to be above water table, allowing it to be imaged without significant diminution due to water content.

Reflections are caused when the transmitted EM wave encounters a contact between two media of differing $\varepsilon_r$. The *reflection coefficient* or *reflectivity* $R$ is related purely to the velocity contrast of the media:

$$R = (V_1 - V_2) / (V_1 + V_2)$$

(5),
where $V_1$ represents the material above the contact and $V_2$ that below. This controls the amplitude (including the polarity) of the returning wave $A_r$, relative to the outgoing wave amplitude $A_0$:

$$A_r = A_0 \times R \quad (6).$$

Either velocity may be higher: *e.g.*, if the EM waves encounter transition into a layer of less compact (more air content) material, such as from a compact sand to a gravelly fluvial deposit, they will travel incrementally faster and produce a reflection of negative polarity (negative amplitude by (5) and (6)); in general, however, waves encounter more compact and hence electromagnetically slower materials with depth, leading to reflections of positive polarity. Typical surfaces that produce high-amplitude (“bright”) reflections include water table, erosional surfaces, and some internal bedding planes (*Beres et al.*, 1999).

While the receiver continues recording for a predetermined time interval, it transmits data *via* a control unit to a laptop computer running RAMAC *GroundVision* data collection software, where the data are automatically saved to a file. The time record of this measurement is displayed on-screen as a single vertical line referred to as a *trace*. The downward-positive vertical axis displays the amount of two-way travel time. Small differences in amplitude caused by differences in $R$ are represented in a trace as “wiggles” to the left or right (for negative and positive polarity returns, respectively) at the appropriate depth (time) on the trace.

The antennae are then moved and further measurements are taken. In Common-Offset (COff) surveys, the main antenna orientation used in the present study, the transmitter and receiver are moved in tandem along a predetermined line, maintaining a fixed distance between the antennae, while in Common-Midpoint (CMP) surveys, typically used for velocity analysis and more closely estimating the depth of an object, the transmitter and receiver are moved in opposite directions.
while fixing the “center” of the pair above a common point. Measurements are taken at constant intervals according to one of these schemes, the resultant traces placed on-screen to the right of the previous traces to create an x-axis that displays the horizontal distance traveled during the course of the survey. A tightly-spaced (relative to $\lambda$) series of traces creates a skeletal subsurface map, between which the software can then interpolate to create a fuller image. This 2-dimensional graphical output is termed a *radargram* (Fig. 10).

If an actual subsurface feature is present, it will likely cause reflections at multiple measurement points, due to the hemispheric propagation path of the EM wave. This will cause the wiggles from adjacent traces to show the same polarity trend and approximately the same amplitude on the radargram, allowing the software to interpolate between the reflection points which represent the feature. As a flat-lying layer will cause reflections at the same depth along its entire length, it appears as a flat layer on a radargram (“c” on Fig. 10C). Reflections off of a dipping layer, however, having traveled an inclined path, will have a shorter travel path (and thus time) to the receiver than one having reflected vertically off the layer, leading to the appearance on a radargram of a layer dipping at an angle shallower than the actual dip (*Lillie*, 1999, p. 138-141). This anomalous dip is further complicated by the more familiar fact that true (as opposed to apparent) structural dip is imaged only when viewed perfectly along-strike. Finally, single point reflectors (*e.g.*, pipes perpendicular to their length; boulders/cobbles; large tree roots) produce very distinctive reflection patterns. The hemispheric outgoing wave detects the reflector long before it is actually above the object, and the longer raypath followed by the wave makes the object appear much deeper than it is. When the reflector is immediately beneath the midpoint of the antennae, the raypath will be shortest, giving the illusion of an object at its shallowest depth. When the survey is
Figure 10: Summary of radargrams. A: Each Common Offset measurement displayed on-screen as vertical trace with small horizontal “wiggles” indicative of returning wave amplitude. Close juxtaposition of hundreds of traces creates radargram. B, C: Typical interpolated radargram generated by GPR Common Offset survey. Axes are right-positive along-line distance surveyed (in meters) and downward-positive two-way wave travel time (left axis, in nanoseconds), converted to depth (right axis, in meters) assuming uniform V = 8.9 cm/ns. Uniform dark grey area at top (a): null space: traces shifted down to account for topography; bright, continuous reflector (b): ground wave, equivalent to actual surface topography; coherent subsurface layer (c): probable geologic layer; hyperbolic reflectors (d): point reflectors (boulders, trees, posts, etc).
continued afterward, the reflector appears to “sink” again, giving a characteristic hyperbolic reflection shape (“d” in Fig. 10C), the apex of which represents the actual location of the object. In the case of large boulders, the orientations of their faces, the width of their top, and (if thick enough vertically relative to the operative wavelength) reflections from the bottom of the boulder can alter the shape of the hyperbola to some extent. Migration, a post-collection processing tool to be described in an upcoming section, attempts to account for these effects and “collapse” the objects to their proper shape and orientation (Lillie, 1999, p. 138-141).

One of the most important steps in interpreting a radargram is establishing an approximate velocity profile to determine the relationship between the displayed travel time and the depth it represents. This can be accomplished by surveying an object of known depth, but in the Hither Woods the only such objects were located in the northern extreme of the park (sedimentological contacts in an exploration well) or were too shallow (buried pipe) to obtain information about the velocity at the depth of interest. Conducting in-laboratory tests on the electromagnetic properties of a sample of the material being surveyed is also a frequently-used technique to test V directly, but provides only a generalized measurement of a specific sample and should not be extrapolated to represent the entire park, particularly if, as I believe, there are many layers which have not been revealed, much less physically sampled directly (digging through even the upper few meters of cobble-rich till would be quite an accomplishment!). The hyperbolic shape of point reflectors was found to be helpful in some cases: the shape of the hyperbola (particularly the steepness of its flanks) depends on both the reflector depth (which must be accurately known: therefore the start-time of each radargram must be checked by eye to be reasonable and be adjusted in processing if necessary) and, more pertinently, the velocity of the wave traveling into it. At equal depths,
steeper hyperbolae indicate slower material: as the antennae move across the surface, the proportional rate of change of the slant distance to the reflector is higher for a shallow object in a slow medium than for a deeper one in a fast medium. Sandmeier Software’s REFLEXW geophysical signal processing software (referred to as REFLEX) used in post-collection processing has a feature which compares the shape to those of idealized hyperbolae based on specific velocities. Comparison of multiple hyperbolae at various places (particularly in the vertical direction) allowed construction of a good approximation of a 2-D velocity profile in most cases.

Equipment Details

A full suite of GPR equipment was available for the survey (Fig. 11). Use of multiple frequencies of antennae is crucial to such a large-scale project. Initial high-frequency surveys (here the frequency-to-wavelength relationship of (3) is again stressed) may be carried out at sites of interesting topography or other surface characteristics in order to establish the shallow subsurface structure and identify key features for further examination. Following wider high- to medium-frequency surveys to obtain a preliminary idea of the extent of these features, low-frequency surveys (which are, due to the nature of their larger sample spacing, generally much more rapidly obtained than high-frequency surveys) may be carried out to explore for much deeper features of interest. The observations of the shallow surveys can then be placed in a much wider context and possibly related to the findings at depth. A piecework three-dimensional subsurface map may be created with good coverage at a variety of frequencies and line orientations.

Each survey requires the use of a pair of antennae attached by fiber-optic cables (for triggering of the two antennae and return of data) to a central control unit. A distance trigger
Figure 11: Typical GPR Common Offset field surveys.  A: High-frequency (500 MHz) GPR survey: transmitter and receiver mounted in same unit (at bottom: *monostatic*) and pushed on a cart.  B: Medium-frequency (200 MHz) GPR survey: transmitter and receiver dragged via sledge, while cart is pushed with laptop and control unit.  C: Low-frequency (25 MHz) GPR survey at field site of present report: transmitter and receiver are separate units operated by individual field personnel, with control unit and laptop managed by third person.
(odometer, hip chain, or keyboard control) can also be connected via a 9-pin parallel port to the control unit. Output from the control unit is fed into a laptop computer running *GroundVision* through its standard parallel port. Data is automatically displayed onscreen and saved after each trace is taken.

High-frequency antennae available for the survey included 800 and 500 MHz systems. These antennae are small enough to fit in a single box (*monostatic* system), and the shielding of the box eliminates a substantial amount of noise external to the system. These surveys could be carried out quickly by a single operator, as the antennae were mounted at the base of a four-wheeled hand-cart referred to informally as the “cart” (Fig. 11A), and were distance-triggered automatically via a spinning-wheel odometer attached to the left-rear wheel of the cart. The relatively compact nature of the soil surface of most survey lines allowed the wheels to make good frictional contact with the surface, hence the value registered by the odometer was generally accurate within a meter or two of the true length of most 100+ meter-long survey lines, as measured by tape measures. Minor post-collection processing of the data was able to adjust for the differences. Typical penetration and resolution of the 500 MHz system are about 4 and 0.05 m, respectively. High-frequency antennae were used only sparingly due to the large vertical extent of the survey and the cobble-rich nature of the subsurface.

I used 200 and 100 MHz antennae for medium-frequency surveys. Typically, they were used in the initial exploration phase of each survey line to coordinate future surveys and identify features of interest, due to their good balance of penetration and resolution (*Davis and Annan*, 1989; penetration and resolution about 15-20 m and 12-15 cm, respectively). These antennae are too large to be included in a *monostatic* system; therefore, each antenna is mounted with its
battery/electronics unit to a vertical wooden bracket. Horizontal wooden spacer brackets of appropriate lengths can then be attached with screws to the vertical brackets to maintain a constant distance between the antennae in a COff survey (0.6 m for 200 MHz, 1 m for 100 MHz). Typically the cart is utilized to carry the laptop and central control unit and pushed along by one operator, while the antennae are dragged by another operator on a flat, flexible polymer sledge (Fig. 11B). Where narrow trails rule out the more preferable side-by-side orientation of the cart and sledge, they can be run in tandem with the sledge leading the cart. In more unforgiving terrain, such as the cobbly surface of the PLC, a plastic-shielded two-wheeled cart can be used to house the 200 (or more-rarely used 400) MHz antennae, and is dragged in a similar manner. Along-line distance is recorded by a device attached either to an operator’s belt or the frame of the brackets. This device (the “hip chain”) contains a spool of thread, one end of which is staked down or tied to a stationary object prior to beginning each survey. The spool spins as it unravels, functioning in a manner similar to the odometer. While this approach is only an approximation of the actual distance traveled, it can be assumed to be a legitimate proxy due to both the sinuosity of the trail (frequent changes in path orientation prevent single-direction errors from accumulating) and the forgiving nature of the cosine function (even significant deflections produce a net distance very similar to that for a straight line). A more laborious but more accurate method involves one operator wearing a harness upon which are mounted both the laptop and control unit. The antennae can then be maneuvered into position by lifting the bracket/antennae structure and moving it the appropriate distance for each measurement, guided by a tape measure, either by the same operator or a second. Distance is then recorded by the laptop operator triggering the system manually using the keyboard for each measurement.
The low frequency antennae (50 and 25 MHz) require multiple operators due to the sheer size of the antennae (Fig. 11C). The harness-and-manual-movement method is used, and generally one operator is needed for each antenna as well as the harness. Wooden spacer brackets 2 meters long are used for the 50 MHz system, while a 4 m-long fabric cordon attached the vertical brackets is held taut to maintain spacing of the 25 MHz antennae. In extraordinary circumstances, both of these antennae may be rotated with the long axis parallel to the direction of the survey lines (“endfire” orientation) to make use of narrow side trails, particularly if there is a need for cross-lines. While the signal obtained from such an orientation is of somewhat lower quality than the standard orientation (with antennae perpendicular to line direction: “perpendicular broadside:” Fig. 12), any additional data regarding the three-dimensional behavior of the subsurface features of interest may be helpful. Both the 25 and 50 MHz antennae are able to penetrate to 40-50 m depth in the glacial sediments of the field area, and have a resolution of about 2 and 1 m, respectively.

**Data Collection**

Collection of radar data commenced in mid-October of 2003 along the Serpent’s Back Trail, a N-S oriented hiking trail just west of the PLC (Fig. 13). This trail is considerably more sinuous than the PLC, but the central path has been cleared of vegetation by frequent hikers and is wide enough to execute tandem surveys using both the 100 and 200 MHz antennae and a cart survey using the 500 MHz system. Most of the hiking trails in the region displayed a similar well-traveled character. The origin (southern end: all initial data were gathered from south to north) of these lines was taken as the intersection of this trail and an E-W trending trail (Petticoat Hill Trail). Petticoat Hill Trail would be used as the origin of lines subsequently run along the PLC.
Figure 12: Comparison of data quality from typical perpendicular broadside (A, B) and “endfire” (C, D) surveys of PLC (see legend of Fig. 13 for location, demarcated by endfire (“ef”) lines). Orientation of antennae shown schematically at far right, with arrow indicating survey direction. A, C: 50 MHz surveys; B, D: 25 MHz surveys. Perpendicular broadside data clearly give superior data, “banding” present on endfire lines. Lines uncompensated for topography.
Figure 13: Map of survey lines gathered during course of survey. Boxed, numbered regions indicate locations of corresponding figures. Also, locations of geophysical and monitoring wells (blue triangles, red dotted circle) referred to in text. Coastal exposure of Figure 7 marked by orange cross. Legend to right indicates extent of each antenna collected on PLC: black lines indicate individual lines assembled to create composite line. “ef” = endfire.
Approximately 340 meters of radar data were collected using each of the three aforementioned frequencies. Cross lines were also collected using the 200 MHz system in tandem orientation along Petticoat Hill Trail and a similar E-W hiking trail which intercepted the two N-S trails at about 120 m to the north (Ocean View Trail) (Fig. 13), totaling about 130-165 m of radar data for each trail. The northern trail was run west-east, while the southern was run east-west. In addition, short (no more than 6 m) 500 MHz lines were collected perpendicular to the trend of the PLC in order to establish the presence of the buried underground pipe described below (Fig. 14).

After initial processing of the data collected along the Serpent’s Back and perpendicular trails, the results were deemed significant and worthy of further investigation using low-frequency antennae. In the course of four roughly monthly expeditions to the study site beginning in May 2004, 1,230 m of the Power Line Cut were surveyed with both the 25 and 50 MHz antennae (Fig. 13), making use of the width of the trail to maneuver the large antennae. The lines stretched from the intersection with Petticoat Hill Trail in the south to the southern margin of the dendritic valley network in the north (Fig. 5B, C, 13). As an initial investigation of the valley network, the 50 MHz survey was extended across this feature by Mitchell Cangelosi for an additional 430 m, for a total of 1,660 continuous meters of data. The 50 MHz antennae were also utilized in endfire configuration to resurvey the previous cross lines in the south of the Hither Woods for deeper penetration.

I acquired data from a nearby geophysical well in mid 2004, courtesy of the U. S. Geological Survey Water Resources Division office in Coram, NY, allowing me to directly compare my preliminary results with a known stratigraphic column and thereby calibrate the radar velocity. To perform this calibration, in October 2004, a N-S trail (the northernmost extent of the Serpent’s Back Trail) which crossed perpendicular to the well 8.3 m to its east was surveyed for 100 m
Figure 14: 500 MHz radargram run perpendicular to trend of PLC reveals hyperbolic reflection caused by buried pipe at 0.9 m depth. Note that the reflections beneath the pipe are disturbed due to energy diffracted from and reflected off of the pipe. Vertical exaggeration ~ 4:2:1.
beginning at the northern intersection with Old Tar Road, and a short (28 m) E-W trail connecting
the N-S trail to the Power Line Cut was also surveyed (Fig. 13). Both trails were surveyed with
200 and 50 MHz antennae, as the vegetation was spread more widely along the path than farther
south and allowed the wider antennae to pass. As stated previously, this provides only an estimate
of the velocity at the northern end of the park, but may be useful elsewhere since a single till unit
appeared to be dominant throughout the park, including near the well.

In January 2005, 50 MHz endfire surveys along the Serpent’s Back Trail were performed
for direct comparison and correlation with previous data acquired on the PLC. Further cross
lines to the east of the 615 m mark of the PLC and to the west of the 220 m mark of the Serpent’s
Back Trail were run with the 100 MHz and 500 MHz systems respectively along valleys between
ridges, in March 2005, to gain information about the along-strike subsurface behavior of the
ridges. 100 MHz data were also collected along the PLC in the same month to increase resolution
in the area, running from the 430-600 m mark. In May, 2005, 100 MHz data were also taken
along the southernmost 200 meters of the PLC, the previously-surveyed portion of the Ocean
View Trail, and a 130-meter north-northwest-trending extension of the latter trail (Fig 13).

I conducted two important calibration surveys in March 2005. A 50-meter-long Common-
Midpoint (CMP) survey was conducted along a planar, N-dipping segment near the south end
of the Power Line Cut to verify the velocity profile calculated from the near-well surveys farther to
the north. Also, a 50 MHz survey was run from north to south (opposite the usual direction)
starting at the PLC 600-meter mark, to test whether there was any variation in radar signal based
on the direction the survey was run (see Meschede et al., 1997). A direction-dependent noise
had been previously observed during the course of an exploratory survey of a field on the Stony
Brook campus, apparently due to the electrical interference of a train station (Fig. 15A, B), but no such dependence was observed along the PLC, (Fig. 15C, D) despite the possible presence of an electrical source in the shallow subsurface pipe.

The final lines were collected in early July 2005 in the northern portion of the study area. Three 50 MHz lines were run to the north of the PLC in order to correlate the hills to those north of the Long Island Railroad track and near the prominent coastal exposure, to trace a feature seen along the main PLC lines (Fig. 13). These totalled about 560 m on secondary trails.

**Geophysical Signal Processing Techniques**

During collection of the data, the RAMAC *GroundVision* data acquisition software applied a number of basic signal processing techniques to improve the display. The most important involved a running-average “dewow” (known in the *GroundVision* software as a DC) filter. This filter removes low-frequency drifts that would otherwise cause strongly anomalous (“wowed”) values after gaining. The result is a much cleaner, brighter display, particularly at depth.

Also applied to the data during acquisition to enhance the display was one of a variety of gain features, which are customized processing steps intended to artificially approximate the original signal strength at depth, much of it having been lost by scattering, attenuation, and geometric spreading of the waves during subsurface travel. The *GroundVision* software applies automatic gain control (AGC), which attempts to create an equal apparent distribution of amplitude throughout the trace. By calculating a vertical average of the amplitudes throughout the entire trace, each sample is scaled according to the mean amplitude (or lack thereof) over a specified window of samples centered on the one to be scaled. Although this method of gaining is valuable for a first
**Figure 15: Signal noise due to direction-dependent energy.**  

A, B: 200 MHz survey on Stony Brook campus displaying difference in noise between NE-SW (top, flipped horizontally) and SW-NE (bottom) surveys, perhaps due to nearby railroad station. Lines taken 1 m apart. Radargrams uncorrected for negligible topography. Vertical exaggeration ~ 6:1.  

C, D: 50 MHz radargam from survey on PLC run S-N (top) and N-S (bottom, flipped horizontally), highlighting minor differences in detail. Some could be changes in the subsurface hydrology from precipitation and percolation (surveys were months apart). Vertical exaggeration ~ 2:1.
approximation, it significantly scales down regions near exceptionally bright and/or shallow reflectors, with the result of the loss of some detail (potentially including important objects) near them. Therefore I typically choose a more specific gain feature during post-acquisition processing.

More sophisticated and varied techniques are available after data collection, using REFLEX. Some of the most important processing techniques applied to the data are various frequency filtering processes, in which only a limited range of frequencies, typically near the peak frequency of the antennae, are preserved as part of the signal. Although stacking data during collection suppresses the amount of random noise recorded during the survey, a significant frequency-dependent noise may still be present, particularly if a strong electromagnetic source is nearby (including electrical lines, cellular phones, or even the high-frequency GPR unit itself). This noise can be removed via a bandpass filter, which Fourier transforms the entire signal into the frequency domain and removes frequencies lower and higher than a selected range. Bandpass filtering is particularly effective at removing high frequency noise, but due to the finite nature of Fourier sine approximation it can introduce sinusoidal artifacts throughout the radargram. The opposite approach (cutting out a specific frequency range) can be achieved using a notch filter, which is useful when noise occupies a specific frequency band and obscures important data. An alternative means of achieving this goal involves a mean filter, which essentially smoothes the signal by averaging a small, user-defined number of samples above and below each sample and sets the observed sample to the average, removing small-scale frequency spikes.

Consistent horizontal noise caused by a constant phase source can be removed by the subtract average filter, which subtracts from each sample the average of a set number of samples to either horizontal side. A similar filter, background removal, removes the average of a user-
selected horizontal band. Performing one of these steps before the gain is applied allows for good results, as a subsequent gain can compensate for the weakened signal.

*REFLEX* also offers a more complete suite of gain options than *GroundVision*. Options include AGC, manually-entered gain schemes in the horizontal or vertical directions, a combination linear-exponential time-varying gain function, or the application of a preprogrammed energy decay curve. I applied manual gain most frequently, as it allowed me to customize the gain for each line and correct for a variety of circumstances which did not vary smoothly or predictably during collection (e.g., small puddles in the survey path, battery weakening, significant change in sediment type). Various other processing steps involve: removing the earth’s “natural” response curve from the signal to isolate the source wavelet (deconvolution: a borrowed technique from seismology. See *Turner*, 1994); vertically realigning traces or the entire radargram so that the start-time of the signal is at its appropriate point (crucial for determining the velocity of the material as described above); correcting for topographic features (described below); assembling and/or extracting portions of lines for further analysis; and stretching lines to their true lengths (important in cases where the terrain, for various reasons, causes an apparent rather than actual survey line length to be recorded: *e.g.*, if a loose patch of soil prevented the odometer wheel on the cart from spinning an appropriate amount).

A final important processing step is known as migration, and is also adapted from seismology. It attempts to account for the fact that the radar signal does not travel outward in a straight line, but in fact as a hemisphere, causing characteristic patterns such as the hyperbolae and spuriously dipped lines mentioned above. Migration attempts to return these patterns to their true form by back-projecting the data to its “true source” along the reflector (*Lillie*, 1999, p. 138-141).
However, as I will explain below, the method requires the surface of the survey to be quite planar and the velocity variations to be simple for it to be possible to migrate the lines effectively.

*Complications Regarding the Survey*

GPR systems are very sensitive to metal objects, particularly shallow ones. Short cross lines perpendicular to the trend of the PLC using the 500 MHz antenna revealed the presence of the pipe buried at about one meter depth there, which caused a significant reflection and disturbed the subsurface representation beneath it (Fig. 14). This ruled out the possibility of using high-frequency antennae directly over the Power Line Cut.

In addition, during the laying of the pipe, the original bedding was undoubtedly disturbed along the entirety of the PLC to a depth of at least one meter (determined from the aforementioned short crosses). Boulders could be frequently seen along the side of the path, some with very little moss or weathering features, suggesting that they had been recently excavated. Several hills are also covered with an odd cobble or pebble pavement, a characteristic not seen frequently on the adjacent NNE-trending hiking trail about 20-40 meters to the west, suggesting that the pipe had been covered by backfill of the excavated sediments. Under such circumstances, any sedimentological analysis involving of the first few meters of depth of hills along the PLC must be viewed with skepticism. This also gave further incentive to avoid relying upon the use of high-frequency antennae, whose signal would be dominated by the disturbed shallow sediments, along the PLC.

The topography and curvature of the hills also creates complications in the survey, which is ironic given that the ridge-and-valley topography is the very reason this area was chosen for survey. The radar system cannot gauge the topography it travels over: any topography traveled
over in the course of a survey is neglected, and the radar output is shown with a flat surface (Fig. 16A). As a result, the raw data includes the depiction of flat-lying layers beneath hills as “smiles,” having been “pushed down” by the “flattening effect,” while layers following topography appear close to horizontal on an uncorrected radargram. This effect can largely be accounted for by taking topographic field measurements and constructing a correction file to be inserted into REFLEX to produce a two-dimensional profile of the surveyed area (Fig. 16B): the zero-time position of each trace is moved downward according to the topographic measurement and user-defined EM wave velocity. Data in this report will be presented in their topographically corrected form, except where otherwise noted.

However, the curvature itself is more difficult to accommodate for. Due to the fact that a survey of a point object beneath convex upward topography will remain anomalously close to the object for an extended period, the hyperbolic shape produced by the object as described above will be distorted into a wider shape: in the extreme case of perfectly semicircular topography with the object at its exact center, the object would appear as a straight line. Conversely, a point beneath concave upward topography will appear as a sharper hyperbola. The curvature also makes the use of migration, which requires the assumption of planar topography, difficult or impossible except over relatively long, planar stretches of the study area. Extensive planar areas, of course, are few and far between along the hilly hiking trails and PLC of Hither Woods, severely limiting the usefulness of migration in most parts of this study. However, the perpendicular cross-lines, which were typically run in valleys or along the shallow sides of hills parallel to their strike, were generally more successfully migrated. However, the subsurface curvature effect was not (and perhaps cannot be) adequately compensated for, in my opinion.
Figure 16: Compensation for surface topography with REFLEX. 500 MHz line from Serpent’s Back Trail from 286-316 m. A: Radargram without topographic correction. Vertical exaggeration ~ 8.3:1 for emphasis. B: Same radargram after application of topographic correction. Vertical exaggeration ~ 2.2:1. “Smiles” in A “pushed” down by assuming flat surface, returned to original orientation in B.
Another complication that faces the survey involves the clast-rich glacial material itself. The presence of multiple scattering objects (almost undoubtedly cobbles and boulders) throughout the subsurface created a chaotic reflection pattern that at times obscured the deeper portions of the radargram (in the manner of Fig. 14), and it was often difficult to establish which linear features were caused from diffraction hyperbolae and which from authentic dipping layers. However, the REFLEX hyperbolic fitting tool typically used for velocity analysis was utilized to differentiate them by comparing the exact shapes. Furthermore, when hyperbolae could be definitively identified, I was able to make use of their flanks to constrain the velocity in the vicinity of the objects. By measuring several of these hyperbolae at various depths throughout the radargram, a preliminary velocity distribution could be determined for the region, and an average velocity chosen for such processing steps as migration and the basic time-to-depth conversion.
The processed 50 MHz radargram from the survey along the Power Line Cut is included in Figure 17 (back pocket), in uninterpreted and interpreted form, with various excerpts displayed in separate figures within the text. This line will be the keystone of the upcoming discussion of radar results, and segments of it denoted by along-line distance will be frequently discussed.

The shallowest feature in the GPR surveys is the ground wave, the band running across the top of each radargram, representing the waveform of the first signal arrival (“b” in Fig. 10C). It is essentially a proxy for topography in topographically-corrected radargrams. The deepest consistently visible feature on the 25 and 50 MHz surveys was a coherent but choppy reflector typically 850-900 ns below the ground wave and of negligible thickness (Fig. 17). Its apparent variation in topography and continuity initially led me to speculate that it might represent the upper contact of a geologically significant layer, since a reflector at such great depth would require a profound EM contrast to create a bright reflection. However, upon closer inspection, the reflection appeared to be intimately related to the ground wave, both in time of arrival and amplitude, and was deemed to be an artifact of interference within the GPR electronics. The same “surface echo” has been noted in other studies whenever the acquisition time on this system has been pushed beyond about 800 ns. This artifact was neglected in processing and analysis, and is marked in appropriate radargrams but will not be granted further discussion.

Results and Interpretations
**Folded Layers**

The high-frequency surveys run along the Serpent’s Back Trail revealed the presence of continuous reflective layers in the shallow subsurface, beginning at only 20 ns travel time (0.9 m) (Fig. 18). Prior to insertion of topography, these layers appeared to be very steeply dipping, and could easily be confused with hyperbolae of boulders (Fig. 18A), especially when using the 500 MHz antennae, which create the highest degree of vertical distortion of the antennae I used (in REFLEX, the lines are expanded vertically to fill the screen, with shallow surveys, relative to the length of the survey lines, expanded to a greater extent).

Once the radargrams were corrected for topography, however, the steepness of the lines was greatly reduced, and it was found that while the layers do not perfectly follow the surface topography, there appears to be a good overall correlation between their orientation and that of the surface (Fig. 18B, C). The form of these layers is quite similar to the fold imaged on the same hill by Bernard (1998; Fig. 8). The most likely explanation for this behavior is that the layers and the hill above are genetically related. This is strong evidence that the entire form of the hill is the result of tectonic folding of formerly flat-lying sediments. If the hills were erosional features (fluvial “fosses” of Fuller (1914, p. 47-48); large-scale erosional remnants from subglacial catastrophic floods of Johnson (1999) and Shaw (2002); ice-eroded bedforms), the radargrams would show flat-lying layers truncated and perhaps only slightly deformed at the surface. Recessional (sensu Sirkin (1982), as opposed to readvance) moraines, passive crevasse-fill moraines, classic “hummocky moraine,” (Goldsmith, 1982; Ciner et al., 1999), and other stagnation deposits should show an essentially unsorted collection of cobbles, boulders, sand, and clay materials that
Figure 18: 500 MHz radargram of Serpent’s Back Trail (see Fig. 13 for location) from 180-220 m. Prominent linear reflectors (yellow dashes) indicative of previously flat sedimentary layers pushed into folds by later glacial advance. Compare with Fig. 8. A: Before topographic correction. Note distortion (may be confused with hyperbolae) and apparent dip. Vertical exaggeration ~ 6.25:1. B: After topographic correction. Vertical exaggeration ~ 2.63:1. C: Interpretation of B.
had been passively let down during downwasting: no discernible pattern, such as that seen here, would likely be visible. Clearly, the tectonic explanation best fits the radar data.

The reflective layers are more apparent on the cross lines run along the Petticoat Hill and Ocean View Trails (Fig. 13). These lines were run parallel to the strike of the southernmost hills along the PLC to give a more complete, three-dimensional description of the behavior of the subsurface structure. Before correction for topography, the layers actually appear to dip linearly to the east (towards the hiking paths), and intersect with the surface. After insertion of a topographic correction file, however, their attitude bends to follow loosely the topography, and eventually changes to a very slight apparent westward dip, away from the high points of the PLC and Serpent’s Back Trail (Fig. 19). The surface slope seems to truncate the reflectors at or very near the surface, perhaps with only a recent soil covering them. This suggests that sedimentary erosion, in general, was more important than folding in this direction. When correlated with the more steeply-dipping reflectors found in the north-south trending lines along the Serpent’s Back Trail and subsequent 100 MHz surveys along the PLC, the overall trend of these shallow reflectors appears to be that of planar reflectors bent significantly on their north and south faces, and perhaps tilted into an overall plunge consistent with the westerly plunge of the ridge topography at this location (see Fig. 5C).

In sum, the folding of these beds appears most significant in the north-south direction, which was perpendicular to the margin of a Wisconsinan glacier responsible for the formation this portion of the Ronkonkoma Moraine (Fig. 5A). As suggested by Bernard (1998, p. 67-68), a very likely explanation is that the glacier advanced from the north and proglacially deformed these sediments, apparently due to folding of formerly flat-lying sediments. These sediments might
Figure 19: 200 MHz radargram (top: processed; bottom: interpreted) of cross line along Petticoat Hill Trail (see Fig. 13 for location). Shallow, gently dipping layers (yellow dashed lines) correlate with steeper reflectors in Figure 18 (approximately perpendicular to page at peak of this figure). Layers follow plunge of local topography. Note that hyperbola flanks (red dashed lines) may be confused with dipping layers. Vertical exaggeration ~ 1.4:1.
include pre-Wisconsinan unconsolidated deposits (earlier outwash, marine and/or Cretaceous to Tertiary coastal plain sediments) and/or wind-, glaciofluvially-, or glaciolacustrine-deposited sediments related to the deformational advance itself. Lower-frequency surveys, particularly those utilizing 100 and 50 MHz antennae, show that these layers can tentatively be traced to a depth of at least 300 ns and perhaps to 480 (13.5 to 21.6 meters) (Fig. 20), giving a potential reference frame for the amount of deformation that probably took place.

In several of the valleys along the PLC, particularly from 550-1200 m (Fig. 17, in pocket), the shallow reflective layers appeared to increase their dip significantly, forming a v-shaped pattern characteristic of a syncline (Lillie, 1999, p. 144-146). Often, several overlying, apparently flat layers were present in these valleys and were truncated at the valley edges by the dipping layers (Fig. 21). These flat-lying sediments undoubtedly represent recent basin fill within the valleys, deposited within structural basins formed by folded layers (see yellow dashed lines on Fig. 17, 21) similar to those in the hill of Figure 18. Indeed, a blanket of medium-grained sand and finer sediments could frequently be seen covering the valley bases, creating a visible contrast at their contact with the material on the hill flanks (Fig. 22). A similar sedimentary contact is quite clear in the radargram of Figure 21.

The fact that at least 8 m of fill appears to be present in some structural valleys (550-800 m, Fig. 13), as judged from the radargrams, suggests that the surface relief of at least some of the hills was much more significant when first formed. It is feasible there has been some erosion of the hill peaks as well, although the deposition was likely much more substantial, having come from a variety of locations and sources (both glacial and postglacial). The approximately 15 m of relief between the top (hill peak) and base of the dipping subsurface reflectors (although an unknown
Figure 20: 50 MHz “endfire” radargram (top: processed; bottom: interpreted) of Serpent’s Back Trail from 0-220 m (see Fig. 13 for location; hill on right is same as Fig. 18). Folding from Fig. 18 continues at least to 300 ns (~ 15 m) depth. Also present are flat-lying reflective layer at 600-700 ns depth and artifact at 900-1100 ns. Vertical exaggeration ~ 2.3:1.
Figure 21: 50 MHz radargram (top: processed; bottom: interpreted) of PLC from 690-840 m (see Fig. 13 for location). Flat lying layers near valley center truncated by steeply dipping layers that appear folded in a manner similar to Figure 18. Flat layers interpreted as later valley fill from (glacio)fluvial processes deposited in structural basin caused by folding. Current topography therefore underestimates amount of total deformation. Vertical exaggeration ~ 4:1.
Figure 22: Photograph of valley fill sediments looking north along Power Line Cut around 750 meters. Medium-grained sands likely eroded from hill sides where vegetation has been eroded. Note visible contrast between light-colored valley fill and orange material composing hill sides.
amount of this total is likely due to fluvial erosion), augmented by a conservative estimate of 1 m of hill top erosion, is consistent with the estimate of folding to 13-20 m depth stated previously. Total original fold amplitude in the proximity of 15 m is suggested, which is quite sizeable compared to observed values in the vicinity of modern glaciers. This comparison is to be discussed in the Conclusions section. As is now apparent, the considerable present surface topography of Hither Hills is actually understating the amount of structural deformation that occurred during the formation of the area, due to subsequent erosion of peaks and deposition in valleys. Such substantial relief must have been amazing to see, let alone hike along!

Deep Reflective Zone (DRZ)

At greater depths, a bright reflective zone was discovered at depths from 400 to 500 ns (18-22.5 m) (Fig. 20, 23). This zone (referred to as the “Deep Reflective Zone,” or DRZ) can be traced discontinuously along the entire length of the surveyed portion of the PLC (Fig. 17), a distance of over 1.6 km, as well as along the 50 MHz cross lines, and appears to be no more than one to two meters thick.

Notable about the DRZ is that it appears even at first glance to be very flat relative to the topography (see Fig. 17). It should be mentioned that the EM velocity of the medium is a critical parameter in determining the depth and particularly the dip of this (or, in fact, any) reflector. Overestimating V will make the layer appear to dip more steeply than it actually does and be deeper, and vice versa. Also, if the surface and DRZ topographies are independent, altering the assumed velocity would change the apparent topographic relationship of the two features in an
Figure 23: Detailed survey of shallow instance of DRZ in trunk of dendritic valley network (see Fig. 13 for location).  **TOP:** Topographically uncorrected 200 MHz radargram from survey of M. Cangelosi.  DRZ as “frowned” layer of hyperbolae that appears randomly distributed.  Vertical exaggeration ~ 6:1.  **MIDDLE:** Same radargram after topographic correction.  Hyperbolae now appear to fall on constant-altitude zone.  Vertical exaggeration ~ 2.85:1.  **BOTTOM:** 50 MHz radargram from Cangelosi study of same area.  50 MHz radar cannot resolve individual point reflectors, giving appearance of coherent layer.  Vertical exaggeration about 2.3:1.
unpredictable manner. Therefore, the topographic behavior of the DRZ was analyzed at two end-member velocities based on the typical velocities observed throughout the PLC: an unlikely fast value of 12 cm/ns; and a correspondingly slow 7 cm/ns. These are compared with the preferred value of the region, 9 cm/ns, in Figure 24.

Analysis of the topography of the DRZ shows that it is, indeed, very flat relative to topography. Although there is a regional dip towards the north at about 0.3º, the DRZ wavers only slightly more than 2 m in a root-mean-squared sense relative to a best-fit plane at this dip for any reasonable choice of velocity (Fig. 24A). In addition, its topography seems to be largely independent of surface topography. The DRZ topography was plotted against that of the surface, and a linear best-fit trend was drawn through the plot points (Fig. 24B). The slope of this plot represents the degree of correlation between the DRZ and surface topography, with a slope of one indicating that the two surfaces are exactly correlated (topographies are parallel) and lower values approaching the asymptotic value of 0 indicating independent topography (DRZ is perfectly flat). Negative values are hypothetically possible if the dips of the topographies are anticorrelated. Even in the extreme case of V = 7 cm/ns, the tested value in which the topographies of the DRZ and surface are most closely related to one another, the best-fit slope is .447: slightly less than half of the surface topography is mirrored in the DRZ (averaged over the entire PLC), and the ground and DRZ surfaces are mostly independent of one another. At V = 12 cm/ns, they are even more strongly independent, with a correlation slope of .053. For the preferred velocity (based upon hyperbola-fitting) of 9 cm/ns, the correlation slope is 0.290. In other words, less than 3/10 of the reflector topography perfectly matches that of the surface, and the processes responsible for the surface deformation predominantly occurred above the DRZ.
**Figure 24: Topographic behavior of deep reflective zone (DRZ) along PLC**, as determined by point analysis of 50 MHz radargrams. **A**: Altitude of reflector with respect to arbitrary datum. Assumed radar velocity has significant effect on dip of non-horizontal surfaces: Here V set to typical value of 9 cm/ns. Trend line automatically assigned: small-scale variations superimposed upon regional dip to N. RMS variation from this dip 2.077 m. **B**: Comparison of reflector depth and surface topography of PLC, using end-member and preferred velocities listed in text and legend. Trend-lines automatically assigned. Linear 1:1 correlation would indicate perfectly parallel topography, more scatter or smaller positive slope progressively less related. Slight correlation between topographies at 7 cm/ns, almost none at 12 cm/ns. See text for details.
On the 25 and 50 MHz radargrams, the DRZ appears to be a coherent layer (Fig. 23C, 25C). The presence of a layer on a radargram could be explained in any of several ways, including: a geologic contact between two units of differing EM properties; a thin layer of differing EM properties within the same unit, e.g. a clay-rich member of a till, or a layer of pebbles or other particles too small to be resolved individually at the 50 MHz wavelength (assuming a velocity of 9 cm/ns, the smallest typically resolvable feature would be about 45 cm); an abnormally elevated (perched) water table (the lowest surface point of the PLC, within the center of the dendritic valley network, is roughly 7 meters above sea level, and the reflector is 2 meters below the surface [as judged by 200 MHz surveys in the same location], placing the approximate minimum altitude of the reflector at 5 m above sea level); or perhaps an anomalous signal due to electronic interference.

Some of these possibilities can be discarded almost immediately. Water is a very good reflector due to its substantial $\varepsilon_r$ (Davis and Annan, 1989). Water table of any kind, whether perched or at sea level, generally provides an unambiguous reflector (even without application of gain in some cases). However, the discontinuous reflector I observed often dropped out and could not be located at several places (e.g. 120-200, 300-420, 590-680, 1060-1130 m on Fig. 17), although a tentative correlation could be drawn as the DRZ on either side of the dropouts seemed to be at the same approximate altitude. Water table should provide a continuous strong reflector, as it represents a potentiometric surface, and while there are some locations (particularly 1250-1650 m of Fig. 17) where this is the case, the overall slightly discontinuous appearance of this reflector is inconsistent with water table. The small-scale variations in altitude are also inconsistent with the planar (or at most only very gradually varying) nature of water table.
The dropouts also probably rule out the possibility of electronic interference. Horizontal noise, like water table, is generally continuous throughout a profile. It also is generally perfectly horizontal, making it easy to remove via post-collection processing steps such as subtracting the horizontal average of each line. However, the DRZ wavers vertically just enough, even after compensation for topography, to avoid being removed by such steps, and does not have the garbled appearance of typical systematic noise. Rather, the DRZ appears to show some patterns characteristic of actual features, such as hyperbola flanks (220, 260, 500, 700, 1530 m of Fig. 17) and the outward appearance of a dipping structure (450, 690-730 m). Therefore, the DRZ undoubtedly is not an electronic interference signature.

The remaining possible explanations involve actual geologic features. More detailed GPR surveys support the idea of a geological source. A small scale GPR survey was carried out by Mitchell Cangelosi in the area of the dendritic valley network in a study designed to explore the possibility that the network originated as a tunnel valley eroded by highly pressurized subglacial meltwater into the underlying, soft sediments (Fig. 13). The surveys revealed that a significant amount of cover sediments had indeed been eroded from the surface (M. Cangelosi, personal communication, 2005). A serendipitous circumstance of this erosion is that the DRZ is anomalously close to the surface at the base of the valley formed in the trunk stream of this tunnel valley (Fig. 17, 23). A 200 MHz survey was then practical at these depths to explore the DRZ in greater detail.

The 200 MHz survey revealed a completely different character of the DRZ, in part because the layer was paradoxically more difficult to resolve at this higher frequency (and, thus, higher resolution). Since the topography of the valley is initially ignored, as described in an earlier section,
the sides of the valley in the uncorrected radargram are forced down during collection, and the truly flat-lying DRZ is forced into a “frown.” This makes it difficult to formulate a proper manual gain scheme that appropriately scales the amplitudes of the reflectors at different depths (times). Therefore, when an initial gain scheme was applied but before topography was inserted, the DRZ, which appears at 200 MHz as an assortment of seemingly randomly distributed hyperbolae, was not clearly visible (Fig. 23A). However, after insertion of topography, the hyperbolae appeared to fall on a constant-altitude layer (Fig. 23B), matching exactly with the anticipated location of the DRZ as judged from the 50 MHz radargram. This suggests that the DRZ, at least at this location, is not a coherent layer, such as a clay blanket (Fig. 23C), but in reality is composed of small point reflectors, probably large cobbles to small boulders, that are spaced sufficiently closely to be irresolvable except via medium- to high-frequency antennae. Subsequent studies using 100 MHz antennae showed that the individual reflectors begin to become separated at this scale, placing their maximum size at about 25-45 cm diameter. A number of geologic solutions can then be presented to account for this odd concentration of boulders and cobbles.

Correlation with Coastal Exposures and Other Geophysical Evidence

In order to discriminate between a contact between two separate units and a thin layer within the same unit, it is necessary to compare the geophysical evidence of the DRZ with proven geologic techniques. Here, the coastal exposure explored by Klein & Davis (1999) and revisited in the course of this study is invaluable in aiding the interpretation of the radargrams. The exposure (Fig. 7A-C) shows a number of folds in the lower half of the bluffs, but the folds appear to be truncated by a flat surface, above which is a very clast-rich, flat-lying unit about 0.7-1 m in
thickness. Pebbles are the dominant clast size within the in situ unit, although larger cobbles could be seen on the hill slough, presumably having been eroded from one of the overlying units. This clast-rich layer could be traced continuously across the entire bluff, is approximately 11 m above Napeague Bay and thus sea level, and is 4-5 m below the top of the bluffs. The layer appears to be a form of basal till (Fig. 7C, 25A), possibly caused by the most recent (Woodfordian) glacier overriding material that had already been folded by a (possibly much) older advance. The genesis of this deposit will be explored more fully in the Conclusions section.

Other geological and geophysical evidence is also available for comparison. Gamma ray log data were taken by the U. S. Geological Survey in the northwest region of the Hither Woods (Dotted Circle, well S-70256 in Fig. 13). Gamma ray logs are a proxy for clay content, as clays often contain gamma ray-producing elements in their pores. Prominent spikes were visible in the gamma ray count at a well depth of about 40-43’ (12.2-13.1 m: Fig. 25B), which denote increasing clay content. The clay could represent a portion of the clast-rich layer observed in the cliff exposure, which has a very fine-grained, compact matrix in some areas. The depth of this spike is fairly consistent with the depth of the layer beneath the top of the bluff.

A geological log was also taken in the tunnel valley, just to the northwest of where Flaggy Hole and Serpents’ Back Trails cross (well S-70264 in Fig. 13), during the installation of a monitoring well. The contact between undifferentiated stratified and till deposits and the underlying stratified unit was found at 2’ above sea level, or approximately 23’ (7 m) well depth (Prince, 1986). This is consistent with the 200 MHz survey across the tunnel valley on the PLC and another perpendicular 200 MHz line taken during the course of Mitchell Cangelosi’s survey.
Figure 25: Correlation of geological, gamma ray, and GPR evidence of DRZ.  

A: Photograph of 1 m-thick pebble-rich layer approx. 4-5 m below surface (11 m above sea level).  

B: Gamma log data from NW portion of Hither Woods area, ~300 m south of exposure.  Spikes in data near 40-43' (12-13 m) below surface (7-8 m above sea level) might represent high clay content of pebble-layer. Data courtesy of C. Schubert.  

C: 50 MHz radargram near geophysical well of B (located 8.3 m to W of 22.6 m mark of this line; see Fig. 13 for location) showing DRZ at ~300 ns (13.5 m) depth (7 m above sea level). Vertical exaggeration ~ 1.2:1.
which placed the DRZ at 6 m below the surface parallel to the well, 2.5 m above sea level (Fig. 23B).

Gamma ray, geologic, and GPR evidence now seem to point towards a common source (Fig. 25): the DRZ represents a prominent meter-thick clast-rich layer with a clayey matrix, overlying intensely deformed material, as noted in the coastal bluffs. The scale of the clasts noted in the outcrop is in accord with the appearance of a coherent layer on 50 MHz radargrams but individual reflectors on 200 MHz radargrams. Also, the relatively increased clay content surmised from the gamma log data could possibly represent the clay matrix in this layer (see Fig. 27 for tentative correlation).

The coastal exposures did not display the presence of any significant shallow folded layers above the clast-rich layer. There are places where small-scale folding at the surface appears to have affected the topography (Fig. 26), and Figure 17 shows that prominent subsurface folding continues to at least 1650 m length and probably beyond, more than 2/3 the north-south length of the park. However, the material above the clast-rich layer appears to be nothing more than typical soil, clay (possibly loess) and outwash deposited by sedimentary processes. It could be that there were once folded layers at the coastal exposure (and perhaps beyond?) as well, but have since been eroded and/or covered by typical slough deposits. Alternatively, the glacial advance processes responsible for the deformation may have ceased at the north end of the south fork, having been replaced by stagnation and passive retreat. In either case, there are no outcrops known to me which clearly display the prominent subsurface deformation observed in Figures 17, 18 and 21.
Figure 26: Coastal exposure displaying small-scale folding affecting present topography. See Fig. 13 for location. Wavelength of fold about 15+ m.
**Potential Sources of Error**

While this survey has revealed definitive evidence for the deformational origin of the Hither Hills, caution must be taken to ensure that the data have been compiled, processed, and interpreted in a reasonable manner. As many precautions as possible have been taken to guard against error, both human and electronic. It is, however, inevitable that some aspects of the survey are subject to irreconcilable error, oversight, inconsistency, and incompleteness. Therefore, it is important to enumerate those limitations and possible shortcomings.

Of the above-mentioned sources of error, incompleteness is most pertinent to the interpretation of the data. It is simply impossible, due to manpower and time constraints and particularly the amount of vegetation, to completely survey the entire park using the available equipment: therefore, it was necessary to extrapolate the relatively scant observed data to the entire region where reasonable. The similarity of the form of the hills throughout the Hither Woods (Fig. 5B, C) makes this assumption appears relatively sound, but a variety of glaciogenic processes may be present within a small area (e. g., Boulton et al., 1999). Therefore, it is strongly suggested (and will be repeated) that further areas of the Hither Hills be surveyed for completeness, especially in the strike-perpendicular direction where essentially only two lines were able to be run.

Also crucial for the interpretation of the data was the choice of processing steps for each line. Due to the heterogeneous nature of the Hither Woods subsurface, each survey line required
a unique sequence of steps to optimize its interpretation. Development of such an individual sequence inherently biases it towards a particular feature (e. g., by choosing a gain intended to bring out a particular depth range more strongly than others, or moving the start time of lines to obtain a better velocity profile), especially after the discovery of an interesting feature such as the DRZ. However, a strong effort was made to let the processing guide the interpretation, rather than vice versa, and to maintain as unbiased an analysis as possible. Processing steps for each line collected during the course of the survey are included in the Appendix, to illustrate the sensible choice of processing sequences.

Other than the two key factors described above, most of the sources of error were fairly minor, especially since the EM noise of the area is minimal. Other sources of error include: the various complications previously discussed in the section entitled “Complications Regarding the Survey;” observation of apparent rather than structural dip (lines were oriented slightly oblique to bedding, but multiple cross lines minimized this effect in many places); inconsistencies from one line to the next due to the variable subsurface hydrological conditions (surveys were taken at all times of season); minor artifacts created in processing (e. g., bandpass filters create banding; the deepest visible reflection is an interference signal); removal of small amounts of data through processing; and misinterpretation of the depth and dip of features due to the fact that only a single value of V may be input for each radargram. Of these, the variable hydrological conditions and single value of V appear to have had the most influence on my interpretation, and have minimal impact on the first-order results of the survey: the conclusion regarding the presence of folded layers overlying a relatively flat DRZ is still robust.
Geological and Glaciotectonic Interpretation

As described in the Previous Works section, there is a wide range of glaciotectonic processes which might be responsible for the formation of the Hither Hills, including (but not limited to) fold or thrust ridges, basal or marginal squeeze moraines, or englacially ferried material.

The diapiric squeeze moraine system seems unlikely, as neither the surface topography nor the internal structure seem to mirror the environment described by Price (1970) or elsewhere, in which pebbles are oriented perpendicular to the ridge crests and hills show steeper distal slopes. No clear clast fabric was visible at any prominent exposures as yet (although future fabric analysis might reveal a pattern), and the hills do not seem to display any preference for steepness. The freeze-on/rafting explanation is also unlikely. While the DRZ might represent a pavement zone where small clasts were frozen-on while cobbles and boulders were left behind, I have not found such an explanation in the literature, and such pavements are generally described as (glacio)fluvial erosional and/or mechanical boundaries, rather than cryological ones (e.g., Clark, 1991; Hicock, 1991). Also, the continuity of the folded layers, in some places being traceable over many hills, argues against the rafting of large intact blocks of material, unless large portions of the park area represent single blocks which were rafted, perhaps out of the present Block Island Sound area, and deposited en masse atop the DRZ. The scales of the hills and of the park itself make this scenario unlikely. It also cannot explain the substantial shallow folding: rather, undisturbed bedding is expected.

The remaining explanations involve proglacial deformation of sediments. There are several lines of evidence which prefer a scenario in which folding is more prominent than thrusting. The folded layers clearly show that the surficial proglacial material behaved in a ductile manner, suggesting
that any thrusting was either blind (Boyer and Elliott, 1982) or passive, having been generated at the center of the overlying folds as a result of the surface deformation (Dahlstrom, 1970). While, as stated before, no definite conclusion regarding the importance of permafrost is likely to be drawn from this report, the ductile behavior tentatively suggests that the deformed shallow substrate was unfrozen at the time. This may have hindered the development of thrust sheets (see the “Potential Role of Permafrost” section), as observed at the Ranco study site of Meyers et al. (1998), in which the deformation observed in the quarry 70 kilometers to the west but still on the Ronkonkoma Moraine required permafrost.

Secondly, large-scale thrusts beneath the hills, particularly listric thrusts rising from a potential décollement such as the DRZ, were not well-imaged in the radargrams (although due to the fact that the radar would likely be imaging cobbly outwash and till in both the hanging and footwalls, actual thrust features may never be visible). In some places small offsets appeared to be present, but even those locations may be the result of transient noise or a small offset in the start time of one particular trace or a series of traces. Some thrust-like features are more visible (e. g., 430-500, 1050, 1600-1650 m of Fig. 17), but their relations to the hills and the DRZ are inconsistent.

Finally, and most convincingly, the overall topographic trend of the Hither Hills, as noted by Bernard (1998) and in an earlier section herein, displays a dip towards the north (the proximal direction of the Wisconsinan ice sheet; see Fig. 27 for a transect along the PLC). In a thrust-dominated, bulldozer-style glacial advance (Fig. 1C), thrust propagation requires the largest, highest hills to be generated proximally to create sufficient gravitational potential gradient and an adequate wedge taper to permit continued foreland deformation (Davis et al., 1983). The
Figure 27: Schematic profile of Hither Hills geological interpretation (profile extended along red line of Fig. 5C). Subsurface portion determined from PLC radargrams and extrapolated to northern radargrams, wells and coastal exposure (see Fig. 13 for locations). Profile trace along PLC created with Global Mapper program. Radargrams labeled by file name: see Appendix for details. Extreme vertical exaggeration.
northerly dip of the Hither Hills is clearly inconsistent with this model, but can be explained by glacial pushing if the glacier responsible for the deformation either deposited more sediment to the south or simply could not override and/or erode the land to such great extent as farther north due to the thinning of the glacier towards the margin. Both of these explanations, as well as some combination of the two, are deemed feasible, and it is here suggested that the deformation at Hither Hills was predominantly in some form of combined folding and push, rather than thrust, moraines (although the latter cannot be eliminated and might be present locally).

The mechanical role of the DRZ must then be explored. While listric thrusts have not been observed originating from this area, folds require a detachment plane as well. It is tempting to assign the detachment surface to this prominent reflector, due to the fact that the folded surface layers approach it (Fig. 17, 20) and its flatness relative to topography (Fig. 24, 27). However, in the absence of tangible geologic evidence beneath the hills such as the offset of matching layers, shear fabric and other shear features, such a conclusion cannot be verified. I prefer a more conservative view: due to the DRZ’s flatness, the structural décollement must lie at or above it.

More puzzling is the potential role and timing of the tunnel valley in the Hither Hills deformational sequence. The fact that it cuts into the hills shows that it postdates their deformation. It is extremely unlikely that it represents a later glaciation event, as the hills north of this location would be even more substantially eroded than they are, and the likelihood that successive glaciations could reach the same vicinity is relatively low. This explanation cannot yet be ruled out; however, it is far more likely that tunnel valley erosion happened during a small readvance during the process of hill formation. It is possible to form tunnel valleys very near the glacial margin: Gilbert Hanson (personal communication, 2005) notes that a tunnel valley with a substantial current developed
less than a mile from the margin of the Stony Brook Sublobe of the Connecticut Lobe near the
SUNY Stony Brook campus. Therefore, it is possible that the deformation in the Hither Hills
region occurred mostly proglacially but also slightly subglacially; while the hill presently immediately
south of the tunnel valley was being formed subglacially and perhaps others proglacially, the nascent
hill acted as a dam for meltwater, which eventually collected into a tunnel valley which then erupted
in one or several episodes to create the drainage network now visible. In fact, it seems possible
that meltwater could collect and intensify in a subglacial valley between the most recent push
features that were partially overridden, forming the nucleus of the tunnel valley. The hill may then be
eroded shortly after forming.

It is unlikely that a single glacial fold complex, whether sub- or proglacial, could propagate
more than two kilometers into the foreland, no matter how large the glacier: only thrust-dominated
events have been found to propagate so far in the literature (e.g., Krüger, 1985; Croot, 1987;
Boulton et al., 1999). It is far more likely that the Hither Hills were formed by more than one
advance episode. Multiple, independent seasonal advance episodes are practically universal in
glaciers (Hewitt, 1967). “Annual” push moraines may be created when a glacier retreats during
the summer ablation season, deposits outwash and glaciolacustrine sediments in ephemeral proglacial
lakes dammed by the most recent moraine (Matthews et al., 1995), and advances during the
winter while deforming the recently-deposited sediments. Modern glaciers displaying similar
processes have been observed in the Himalaya (Hewitt, 1967), Iceland (Sharp, 1984), Norway
(Worsley, 1974) and Svalbard Island (Huddart and Hambrey, 1996). Similar features apparently
representing repeated (not necessarily annual) later retreat stages of the same Wisconsinan ice
sheet presumed to be responsible for the deformation of the Hither Hills have been documented
throughout Connecticut (Goldsmith, 1982), Massachusetts (Larsen, 1978; Larson, 1982; Oldale and O’Hara, 1984), Block Island (Sirkin, 1976), and Maine (Smith, 1982). Borns (1973) has assembled an excellent summary of the evidence for marginal fluctuations in the New England area. Judging from the morphology of the hills and Figure 13, there appear to be moraines from at least 11 push events in the surveyed area of the PLC.

Fuller (1914, p. 210) mentions that in order to generate sufficient leverage to build a push moraine, modern glaciers appear to require a thickness at least twice the height of the moraines they create. Using this rule of thumb and an estimate of 15-20 meters of deformation, the margin of the glacier during the Hither Hills advance needs to have been at least 30, and perhaps as much as 40, meters thick. This hardly seems thick enough to entail a massive readvance of the entire Wisconsinan ice sheet or even the Rhode Island Lobe (to which Sirkin (1982) attributes the glaciation of the eastern portion of the south fork). Rather, the thickness might correspond to the readvance of a small, isolated ice tongue or sublobe. Indeed, one of the unique features of the Hither Hills is the fact that they are isolated along this particular area of the isthmus: similar features are not seen to the east on the Montauk Peninsula proper, nor to the west in Napeague, nor anywhere on the south fork of Long Island within the area glaciated by the supposed Rhode Island Lobe. I consider this to be evidence that the Hither Hills are a result of the oscillations of a small remnant sublobe of the Rhode Island Lobe of the Wisconsinan ice sheet, while the majority of the lobe retreated in a fairly continuous manner (Fig. 1C, 28).

The most significant factors responsible for the repeated fluctuations this remnant sublobe likely are climatic effects (global or localized) or differences in the topography or substrate which affect the flow of the ice (Boulton, 1986). The first of these will be discussed in the next section.
Figure 28: Proposed genesis of Hither Hills. Pro- and partly subglacial folding of unconsolidated layers forms hills (Fig. 18) while subglacial overriding, deposition and shearing creates flat, pebble-rich deep reflective zone (Fig. 17, 23, 25). Subglacial tunnel valley may develop very near margin during partial overriding, eroding previously-formed hills but not DRZ. Repeated marginal oscillations (see Fig. 1C) create future hills deposited and deformed above DRZ. Modified from Hart and Waller’s (1999) description of proglacial compression and subglacial overriding, sedimentation, and shearing at Melabakkar-Ásbakkar, Iceland. Note general similarity of deformation to profile in Fig. 27.
There are many reports of ancient small glaciers which have been controlled by topography during oscillations, typically advancing farther in valleys. These events were particularly common in the New England area (e.g., Black, 1982; Stone and Peper, 1982; Smith, 1982; Oldale and O’Hara, 1984). Unfortunately, no report known to me has been able to discover a significant submarine channel fitting the dimensions and location of the Hither Woods region. Such a discovery would do much to advance my argument. Perhaps of more import is the substrate geology in the Hither Woods. The presence of a localized pocket of soft proglacial sediment has been postulated to cause a more extended advance of a portion of a glacial margin, as the sediment may deform subglacially and ease the glacier’s movement (Boulton and Jones, 1979). Perhaps there was a small area of extensive, unconsolidated drift or other deposits in the region (such as a glaciodeltaic fan), allowing subglacial deformation and an abnormal advance process. The presence of Gardiners Island almost directly to the north of the study area is a tantalizing piece of the puzzle, as it may tell a portion of the story regarding either or both the topographic and sedimentologic control on the behavior of the glacial margin. Perhaps future geological work in southwestern Block Island Sound and/or Gardiners Island might illuminate the situation more clearly.

In the generalized push-and-fold model, the importance of the DRZ is not entirely clear. It likely represents the surface overridden by the ice sheet responsible for the folding of Hither Hills. However, its relation to the ice sheet itself is more ambiguous. It could represent a basal unit deposited by the same Wisconsinan ice sheet which deformed Hither Hills, e.g. a meltout or lodgment till. It could also represent the erosional remnants of an earlier glaciation, either from a previous Wisconsinan ice sheet (Altonian substage) or perhaps even Illinoian drift. The coastal exposure (Fig. 7) suggests that the deformation beneath the pebble layer which represents the
DRZ is more substantial than that above the pebble layer. *Fuller* (1914) notes that in almost every locality he mapped, the deformation during the Montauk advance (which he assigns to the Illinoian period) was much more substantial than the recent Wisconsinan glacier which formed the Ronkonkoma Moraine. Also, the fact that the folds in the exposure have been cut at the top by the cobble layer shows that the folds predate the Hither Hills deformation (or at the very minimum the deposition of the pebble layer was syntectonic). Therefore, it is my opinion that the DRZ represents either a cobble and boulder pavement which is an erosional remnant of a prior glaciation (likely Altonian, or penultimate, rather than Illinoian, due to the fresh appearance of the folds) that was overridden by the Woodfordian (most recent) glaciation, a basal till pavement related to the Woodfordian itself as it overrode (and perhaps sheared subglacially) the deformed deposits of the prior glaciation, or a combination of the two (Fig. 28).

*Paleoclimatic Implications*

Due to the nature of their formation (in that they require the seasonal deposition of readily-deformed outwash sediments), “annual” push moraines are restricted to warm-based (temperate) glaciers (*Bennett*, 2001). More specifically, moraines are commonly constructed during a transition in climate from cold to warm periods (*Lowell et al.*, 1999), when the seasonal mass balance of the glacier is at a critical juncture between net accumulation and ablation phases and the glacier cannot override and destroy the very moraines it creates. These boundary conditions suggest push moraines in general can be of great significance in climate studies (*Boulton*, 1986), perhaps representing a period of great climatic upheaval. In this respect, the Hither Hills take on new importance.
If the Hither Hills do represent literally annual readvance moraines, they would be among the largest such features to have been described within my knowledge. Most of the annual push moraines described in the literature are only on the scale of a few meters or less tall (e.g., Worsley, 1974; Sharp, 1984; Boulton, 1986). They bear closest resemblance, in morphology, distribution, and amplitude, to moraines in coastal Maine described by Smith (1982). Those features were deemed to represent short-term (perhaps seasonal) but large-scale fluctuations of the Woodfordian tidewater margin. However, the marine limit likely never entirely overran the Hither Hills area (G. Hanson, personal communication, 2005), eliminating a tidewater source. The simplest alternative explanation for the unusual size of the Hither Hills is that the repeated advances responsible for their formation occurred on a longer-than-annual time scale.

To verify that climatic fluctuations are possible on a longer-than-annual time scale, other climatic evidence should be consulted. Ice cores from Greenland have shown that oscillations are possible on an intermediate (millennial) scale, with abrupt (possibly decadal) heating periods followed by slower cooling periods, in a cycle known as a Dansgaard-Oeschger event (Grootes et al., 1993; Dansgaard et al., 1993). Some of these fluctuations have been found in ice cores from both hemispheres (e.g., Thompson et al., 1998), illustrating that global climate change was likely responsible for the fluctuations and can be altered on surprisingly short time scales (Fig. 29). However, the resolution available to even the most recent studies in the Pleistocene does not seem to be sufficient for determining the possibility of repeated decadal or shorter fluctuations of an ice margin, and the millennial cycles currently observed are simply too long to explain the Hither Hills. It appears that climatic verification of my theory will require further advances in climatology and, as Lowell et al. (1999) suggest, the substitution of accurate models for palpable climatic data.
Figure 29: Correlation of globally distributed ice cores. $\delta^{18}O$, oxygen isotope ratio, is a proxy for climate (higher values = warmer). Correlation indicates global rather than local climate change. Distribution of Hither Hills push/fold moraines can be correlated with climate records to establish importance of local environment in retreat processes. Globally correlated fluctuations in general warming trend around 20-15 ky BP thought to represent larger-scale versions of possible climatic forcing function of margin oscillation at Hither Hills. Modified from Thompson et al., 1998.
The scale of the Hither Hills can be used as a crude proxy for estimating the size of the glacier responsible for their formation. As mentioned in the previous section, the ice tongue responsible for the Hither Hills deformation appears to have been between 30 and 50 meters thick. This information could be used in a climate model to judge the localized relative effects of such factors as temperature, precipitation, sediment and topographic control, and possible marine interaction on the Wisconsinan ice sheet.

Also, paleoglacial marginal landforms such as terminal moraines have been used in many previous reports to reconstruct past glacial positions (e.g., in New England, Black, 1982; Stone and Peper, 1982; Smith, 1982; Sirkin, 1998). While the Hither Hills tend to be piecewise, discontinuous patches of hills which reflect a local marginal trend and not a single, continuous, linear chain (Fig. 5B, C, 6), it might be possible, with the help of a wider topographic and radar study, to map the various marginal positions of the sublobe in a manner similar to Figure 2. Armed with this information, it might be possible to form a detailed history of how this particular sublobe retreated from the area, and thus the importance of climate relative to terrain in the retreat process.

Finally, the reader may note that with one exception, I have solely referred to the glacial age of formation of the Hither Hills as Wisconsinan (only specifying the Woodfordian substage as the likely candidate in one recent section) and never as a terminal moraine. There is good evidence to suggest the presence of a more southerly moraine which now lies beneath the Atlantic Ocean, mostly in the form of till and deltaic deposits south of the Ronkonkoma Moraine (Sanders and Merguerian, 1994; Sirkin, 1995, 1998; King et al., 2003; G. Hanson and K. Schmitt, personal communication, 2005). In this case, the farthest extent of the Wisconsinan ice sheet likely surpassed the Hither Hills before its formation, and the Hither Hills portion of the Ronkonkoma Moraine is
not terminal in the literal sense, but rather an active recessional moraine. The role and relative
timeframe of the Hither Hills deformation event might then be a pivotal piece of information
incorporated into a larger-scale history of glacial depositional processes. While it is also highly
unlikely that the Hither Hills have been completely overridden since formation, the idea that an
older-than-Woodfordian glacier is responsible for the Hither Hills cannot be eliminated until dating
techniques such as optically-stimulated luminescence, radiocarbon dating, or perhaps pollen analysis
are applied (provided that suitable materials can be found) and a more concrete determination
made. These ideas are important for properly reconstructing the margin and history of the entire
Wisconsinan glaciation in the New England area.

Regional Hydrological Implications

The populace of Long Island, as mentioned in the Introduction section, is almost entirely
dependent on groundwater for its drinking water. The south fork (particularly the Montauk
peninsula) is isolated from the mainland, which has access to a much greater portion of the underlying
aquifers. As a result, only a shallow lens of freshwater is available in the aquifers beneath the
Montauk area (Colabufo and LaMonica, 1996). Upconing of saltwater and even the presence
of iron have been detected in area wells, showing that the aquifers are nearing the maximum
possible production limit (Prince, 1986; Colabufo and LaMonica, 1996). Therefore, knowledge
of the distribution of aquifers and aquicludes is crucial to the Montauk area.

The presence of folded layers in the Hither Woods area could potentially affect the
distribution pattern of the aquifers in the area by tilting them (and aquicludes) towards or away
from certain recharge basins (Bernard et al., 1998). In addition, the deep cobble layer, potentially
having a clay-rich matrix, could present a significant obstacle to groundwater flow in the vertical direction (note possible clay content in Fig. 25B). In the course of this report, it has been established that the DRZ extends almost completely across the south fork in the Hither Woods area in the north-south direction (Fig. 17, 27). The east-west extent of this till is unknown, although a similar reflective till has been imaged beneath the “Walking Dunes” immediately to the west of the survey region (J. Girardi, personal communication, 2005). For purposes of groundwater modeling, the east-west distribution of the DRZ should be mapped, and a more complete surveying of the folded hills should be carried out: for although there has apparently been sufficient supply of fresh groundwater for the past and current population of the Montauk Peninsula, future needs will require a more complete understanding of the subsurface geometry and the recharge process to make the most use out of the already-stressed aquifers.

General Conclusions and Future Work

In summary, the radar investigations undertaken in this study have revealed a variety of glaciotectonic features in the Hither Woods region. The presence of folded layers at shallow depths (approximately 90-300 ns, or about 4 to 14 m) suggests that significant glacial deformation has taken place to form these hills (Fig. 18). In some places, the scale of this deformation has been masked to a certain extent by subsequent (glacio)fluvial deposition (Fig. 21), but the overall character is quite clear in the radar images. Folding was likely mostly proglacial with some subglacial component, as evidenced by the presence of a tunnel valley partially eroding some hills.

Also imaged was a deeper, fairly continuous reflective zone that is notably flat with respect to sea level (Fig. 17, 27). The décollement responsible for most of the deformation of the overlying
hills must therefore be at or above this zone. It is thought that this reflector represents a pebble layer (observed in a coastal exposure to the north of the study site: Fig. 7, 25, 27), due to its thickness and character. It likely represents the surface overridden (and possibly sheared) by the Woodfordian glacier as it repeatedly advanced and retreated, respectively deforming and depositing fresh outwash, glaciodeltaic, and/or glaciolacustrine sediments (Fig. 29). Each advance, likely on a longer-than-annual time scale, formed a hill or small chain of two or three hills. Preliminary analysis suggests that there are about 11 separate advances in the study area, although further analysis will likely alter this approximation.

Although the first-order interpretation of these features is robust, much work still remains to be done to polish the details of this study. In order to validate many of the conclusions presented herein, and to refine the estimates of the magnitude and number of advances, it is all but requisite to survey a more extensive portion of the Hither Woods region. There may be more significant clues underlying isolated portions of the park, perhaps involving deformation unrelated to that observed by my studies. While an exhaustive survey would be a momentous undertaking, it would certainly answer many geological, climatological, and hydrological questions and perhaps raise many new pertinent ones.

Structural analysis on the nearby coastal exposures or a small artificial cutout in the hills themselves would do much to determine more exactly the nature of deformation in Hither Hills. Pebble fabric, dip, and degree of compaction could all illuminate the dominant process(es) of deformation. In order to confirm that the Woodfordian glacial substage was responsible for the observed deformation, dating techniques such as optically-stimulated luminescence, radiocarbon
dating, and pollen analysis should be employed. Other available geophysical and geological log data should be consulted for ground-truthing.

Finally, if possible, these features should be correlated with other proven glaciological evidence, both in the Long Island area and across the United States and Canada. If it can be determined that similar features in widely spread areas were contemporaneous and have a similar origin, it would suggest that the explanation I have put forward is at least feasible, if not likely.
References


Jakobsen, P. R., Overgaard, T., 2002. *Georadar facies and glaciotectonic structures in ice marginal deposits, northwest Zealand, Denmark.* Quaternary Science Reviews 21, p. 917-927.


