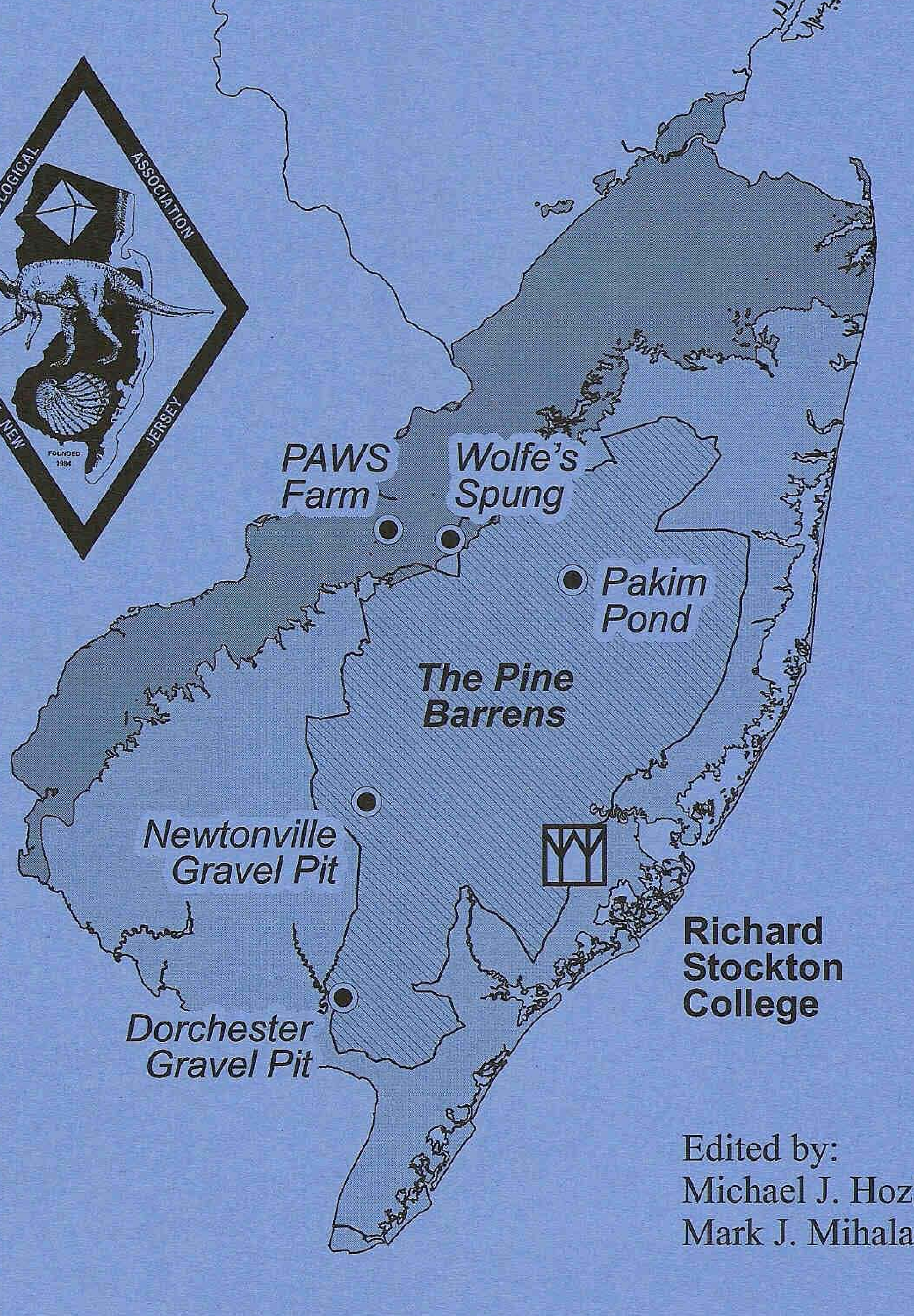
Periglacial Features of Southern New Jersey

Geological Association of New Jersey
20th Annual Meeting
October 10-11, 2003

Field Guide and Proceedings



Edited by:
Michael J. Hozik
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Introduction to the Proceedings Volume of the 20th Annual Meeting of the Geological Association of New Jersey October 10-11, 2003

Michael J. Hozik
The Richard Stockton College of New Jersey

It is hard to believe that this is the 20th Annual Meeting of the Geological Association of New Jersey. As I look back over the history of GANJ, I am struck by the huge debt we owe to Dr. Fredric Goldstein for his vision of an association of New Jersey geologists, and his initiative to get us going. As someone who was present at those early meetings, it is rewarding to see what this organization has become. This year's meeting is a fine demonstration of why GANJ needs to exist and what we can accomplish.

The inspiration for this year's meeting comes from Mark Demitroff, a local natural historian. Mark grew up in Buena Vista Township (locally pronounced **Byōṓ**·na), and spent many of his formative years exploring, fishing, and wading in Pinelands spungs. He knew where to dig to find native American artifacts on their shores. He also collected ventifacts, even though he did not realize what they were. As a student of Dr. Peter Wolfe's at Rutgers, Mark learned the connection between the spungs and South Jersey's polar past. Like Mark, Wolfe grew up on a Pinelands farm and spent his childhood playing among these basins. Gradually, Mark made more mature and careful observations, started reading about arctic landforms, and searched the Internet for more data. As he began to formulate his ideas, he realized the need to talk with someone who was more familiar with periglacial features. After extensive searching online, he connected with Dr. Hugh French, Dean of Sciences, at the University of Ottawa. Mark sent Hugh some of his data, maps, and photos, and Hugh became very interested. He wanted to actually see these features on the ground. Mark proceeded to find funding to through the National Park Service to bring Hugh to South Jersey for a field excursion. We have Julie Akers and the Great Egg Harbor Watershed Association to thank for their support.

Hugh visited the localities Mark had found, and has worked with Mark as a colleague and mentor. Together they have published several papers on periglacial features in southern New Jersey. Their work continues, with minimal financial support, mainly driven by their pure enjoyment of geologic discovery.

About the time Mark was beginning to connect with Hugh, Mark brought his photos and maps to Stockton. Ray Mueller and I encouraged Mark to continue his work, and to think about leading a Geological Association of New Jersey field trip. I am proud to see Mark's and Hugh's work presented at this meeting. Together they have made some important contributions to our understanding of the geology of southern New Jersey. They have raised serious questions about some older interpretations of these features, and they have made some startling suggestions. It is the hope of the GANJ Executive Committee that this meeting will serve as a forum for these ideas to receive the kind of attention they deserve. We hope this will be a lively and stimulating meeting.

Permafrost – Past and Present

Geological Association of New Jersey Keynote Address
The Richard Stockton College of New Jersey
October 10, 2003

Hugh M. French
University of Ottawa

INTRODUCTION

Permafrost refers to perennially frozen earth materials. It is ground that remains at or below 0° Celsius for at least 2 years. It is a temperature condition that reflects a negative heat flow at the ground surface. The best way to illustrate this is a graph of ground temperature against depth (Figure #1). The temperature change with depth is known as the geothermal gradient. If the ground surface temperature is negative, the geothermal gradient means that the ground temperature progressively warms-up with increasing depth. The position where the temperature reaches 0° Celsius marks the bottom of permafrost.

Permafrost can vary in thickness from a few meters to several hundreds of meters.

Delineation of the top of permafrost is more complicated because air, ground and near-surface temperatures fluctuate seasonally. The mean maximum and minimum ground temperatures are plotted on figure # 1. The point at which ground temperatures cease to vary is known as the depth of zero annual amplitude. This is also indicated. The mean annual ground surface temperature is obtained by extrapolation of the geothermal gradient upwards from this point to the ground surface.

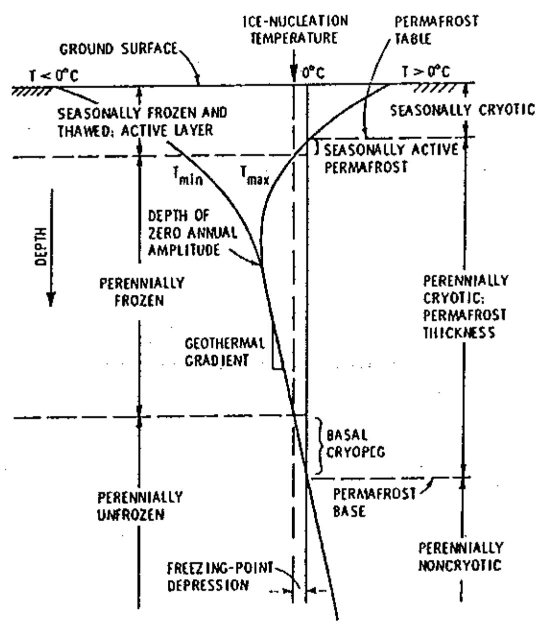

Figure 1. Graph to describe the permafrost temperature: depth relationship and associated terms. From ACGR, 1988, figure.2.

Figure # 1 also identifies the near-surface layer of permafrost that thaws and freezes annually. This is known as the active layer. The depth of the active layer is the point where the maximum ground temperature curve crosses the 0° Celsius line. The active layer varies from less than 0.5m in clay and other unconsolidated sediments to over 6.0m in consolidated bedrocks such as granite or quartzite.

Figure # 2 illustrates the more general thermal picture of the earth-atmosphere interface. In permafrost regions, the mean annual air temperature (MAAT) in winter is usually cooler than the mean annual ground surface temperature (MAGST). In summer, the reverse is true. This difference between air and ground temperatures is referred to as the 'surface offset'. The amount of the surface offset depends upon such factors as the thickness and duration of snow, and the nature of the vegetation cover. Likewise, there is an offset between the mean annual ground surface temperature (MAGST) and the temperature at the top of permafrost (TTOP). This is known as the 'thermal offset'. The latter largely depends upon the thermal conductivity of the earth material in question. This, of course, is also a major determinant of the thickness of the active layer.

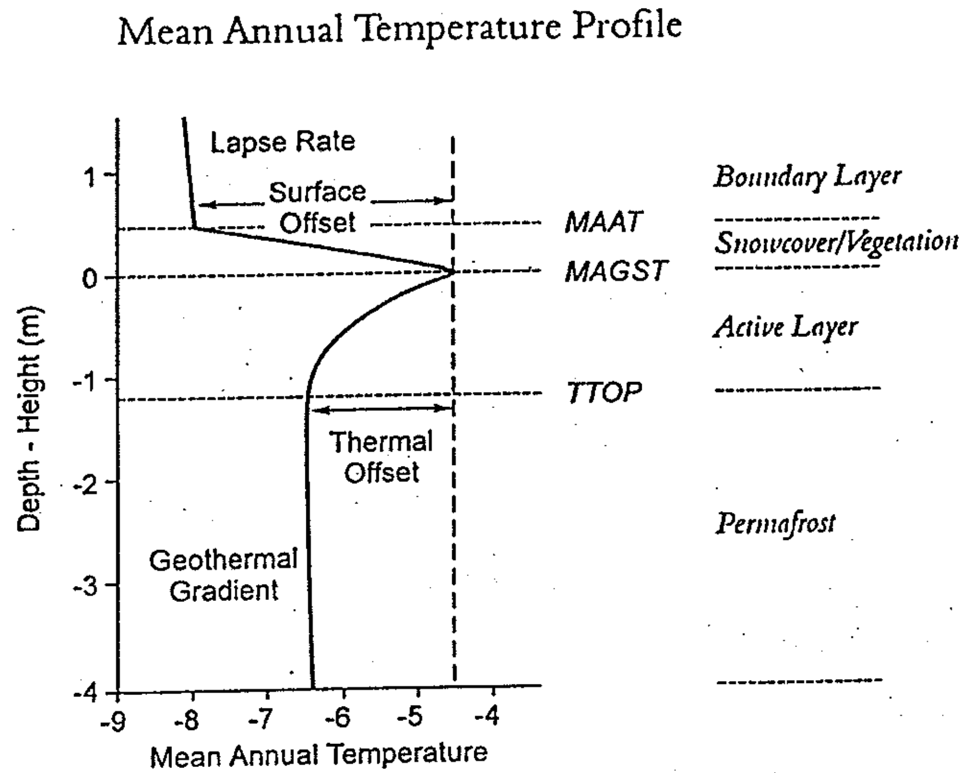

Figure 2. Graph showing the mean annual temperature profile near the ground surface in a permafrost region. From Smith and Riseborough, 2002, figure 3.

PERMAFROST DISTRIBUTION

Permafrost occupies approximately one quarter (25%) of the Earth's land surface (Figure # 3). In the northern hemisphere, permafrost constitutes extensive areas of northern Canada, Alaska and Siberia. It also occurs over wide areas of the Qinghai-Tibet Plateau of

China. Permafrost also occurs in mid-latitude mountain areas such as the higher elevations of Scandinavia, the Rockies of North America, and the European Alps.

For descriptive purposes, permafrost is usually divided into zones of continuous, widespread discontinuous, and scattered discontinuous permafrost where respectively over 80%, 30-80%, and less than 30% of the terrain is underlain by perennially frozen ground. These zones are broadly indicated on this overhead. In the northern hemisphere, the southern boundaries of these zones are defined by the mean annual air temperature isotherms of -5, -4, and 0° Celsius respectively.

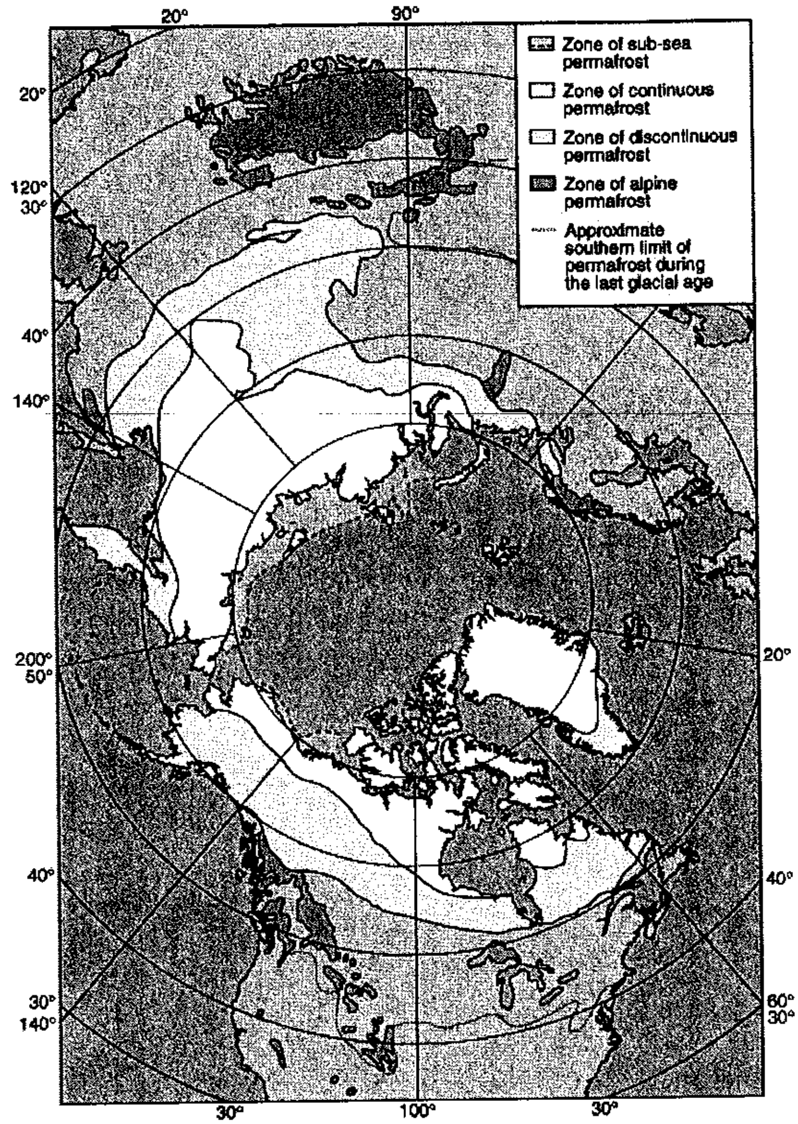

Figure 3. Map showing the distribution of permafrost in the northern hemisphere. Adapted from Péwé, 1991, figure 1.

At the more local scale, the distribution of permafrost reflects a number of subtle and complex factors. These include the thermal conductivity of the substrate in question, the nature and type of vegetation present, the thickness of any organic cover, the thickness and duration of the snow cover, the presence or absence of water bodies, and local drainage conditions. Accordingly, the southern boundary of the discontinuous permafrost zone is sometimes difficult to identify. In the scattered discontinuous zone, permafrost is restricted to 'islands' of peaty materials. In this zone, deep seasonal frost rather than permafrost is the dominant ground thermal regime. Finally, outside of the permafrost regions, such as here in New Jersey, shallow seasonal frost occurs during the winter months.

Thus, there is an entire range of ground thermal regimes ranging from areas of widespread perennially frozen ground (permafrost) through to areas of discontinuous permafrost with deep seasonal frost, and finally, to areas of deep, and then progressively shallower, seasonal frost.

CURRENT GEOTECHNICAL IMPORTANCE

The importance of permafrost often relates to the presence of ice, usually near its melting point, within the ground. Characteristically, the uppermost layers of permafrost are ice-rich. To understand why, soil physics needs to be employed.

In soil and rock that is subject to freezing, water moved in response to a thermal gradient, from warm to cold. This capillarity process is called cryosuction. Depending upon the speed at which freezing occurs, water either freezes in places, to form pore ice, or migrates towards the freezing plane to form lenses of intrasedimental ice. The latter forms either by ice segregation during permafrost development or by intrusion of water under pressure into frozen ground.

Figure # 4 diagrammatically illustrates the ice distribution pattern that typically exists when seasonal freezing of the ground occurs from the surface downwards. The cryofront is the downward-propagating 0° Celsius temperature line. The freezing front is some distance behind this because percolating groundwater contains minerals and has a depressed freezing point. The separation between the freezing front (i.e. the temperature at which individual ice crystals form to produce pore ice) and the occurrence of the lowest ice lens is termed 'the frozen fringe'. In this diagram, the thickness of ice lenses decrease progressively towards the surface reflecting the fact that the downward advance of the freezing front decreases exponentially.

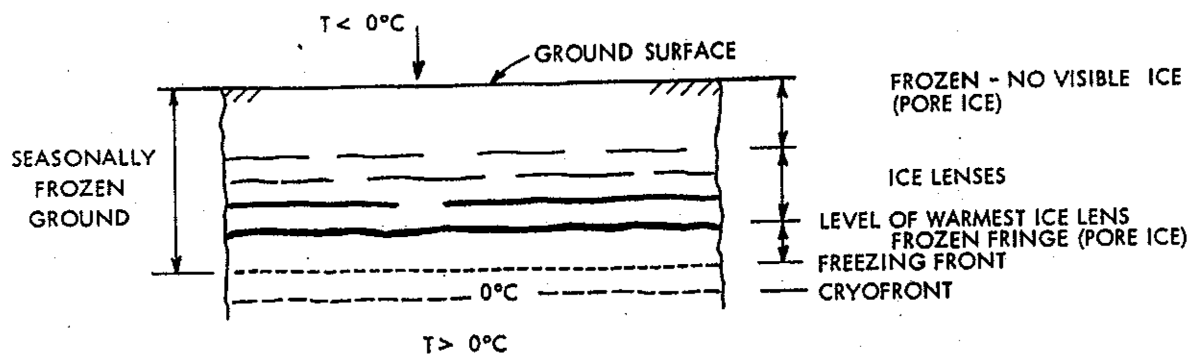

Figure 4. Diagram showing the relative positions of the frozen fringe, the freezing front and the cryofront during freezing of a fine-grained frost-susceptible soil. From ACGR, 1988, figure 5.

If we now assume that this freezing process occurs in an area where permafrost already exists, the picture becomes more complicated. It is well known that moisture migration can occur in frozen sediments in response to a similar temperature gradient, that is, from warm to cold. Thus, in winter, the ground surface is colder than the permafrost beneath and in summer the reverse is the case. Therefore, there is an upward movement of water towards the base of the seasonally frozen layer in winter and a downward movement of moisture towards the permafrost table in summer. The result is the formation of an ice-rich zone at the base of the active layer and at the top of permafrost.

Figure # 5 provides data that illustrates what actually happens. It shows the volumetric water content profile obtained from a glacio-lacustrine sediment near Mayo, Yukon

Territory, Canada. Ice contents in excess of 50% by volume characterize the top 0.5 to 2.25m of permafrost.

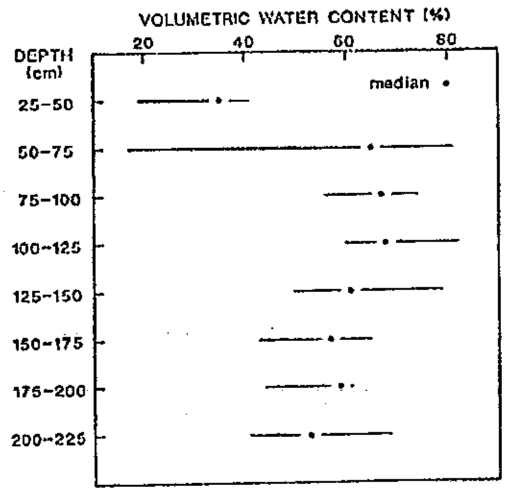

Figure 5. Graph showing the volumetric water content profile of glaciolacustrine sediments at Mayo, Yukon Territory. The profile shows minimum, median and maximum values of water content in core samples collected from six holes, continuously sampled over 25 cm intervals. From Haeberli and Burn, 2002, figure 9.11.

The volume expansion of 9% that accompanies the phase change of water to ice causes an upward heave of the ground. This is further amplified if lenses of intrasedimental or segregated ice form. The result is the frost heave that all North Americans outside of southern California and Florida know about! It results in potholes in roads, damaged foundations to building, and millions of dollars of damage annually. In most instances, there is little that one can do about frost heave because the cost of preventive measures is exorbitant and justified only in certain circumstances.

The thermal condition that permafrost represents, and the ice that is contained within the permafrost, are of fundamental importance. For example, Figure # 6 illustrates the thaw-instability of permafrost. An increased mean annual temperature at the surface causes a shift from thick permafrost and a thin active layer to thinner permafrost and a thicker active layer.

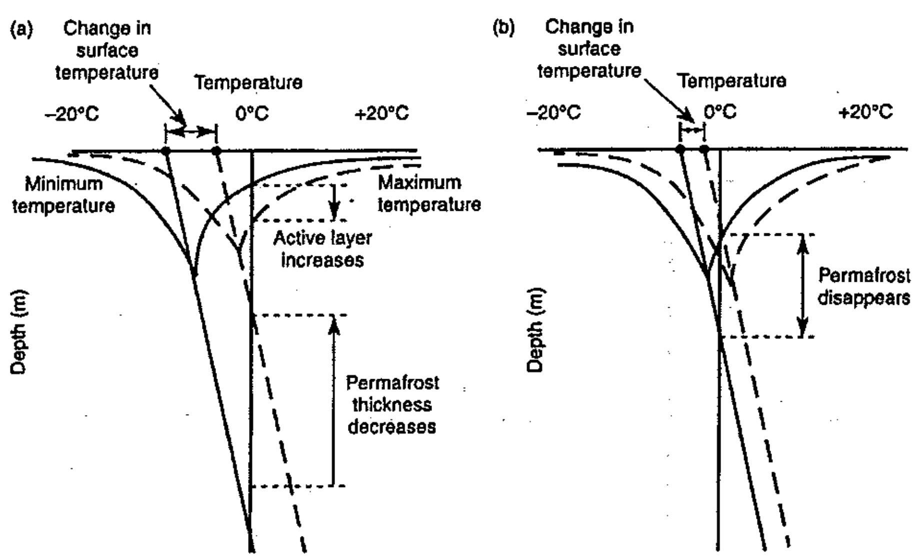

Figure 6. Graphs showing equilibrium ground temperature profiles following a climatic warming of 4°C in (a) continuous permafrost zone and (b) discontinuous permafrost zone. From French, 1996, figure 17.3.

Today, there is much concern about global climate warming. Since permafrost is a climatic phenomenon, climatic changes lead to an altered permafrost regime. For example, much of Europe has just experienced one of the warmest summers on record. There is concern that permafrost in the high mountains will thaw. If the ice that fills the cracks and interstices of bedrock melts, the rock may weaken and cause the failure of structures constructed at high elevations. This figure (Figure # 7) shows recent drilling to investigate permafrost conditions in Switzerland, where there is concern for the stability of high-altitude cable car stations and other ski-related facilities. Likewise, in the high-latitude northern regions of Canada, Alaska, Siberia and elsewhere, there are concerns that permafrost will thaw, causing widespread damage to buildings, roads and other structures.

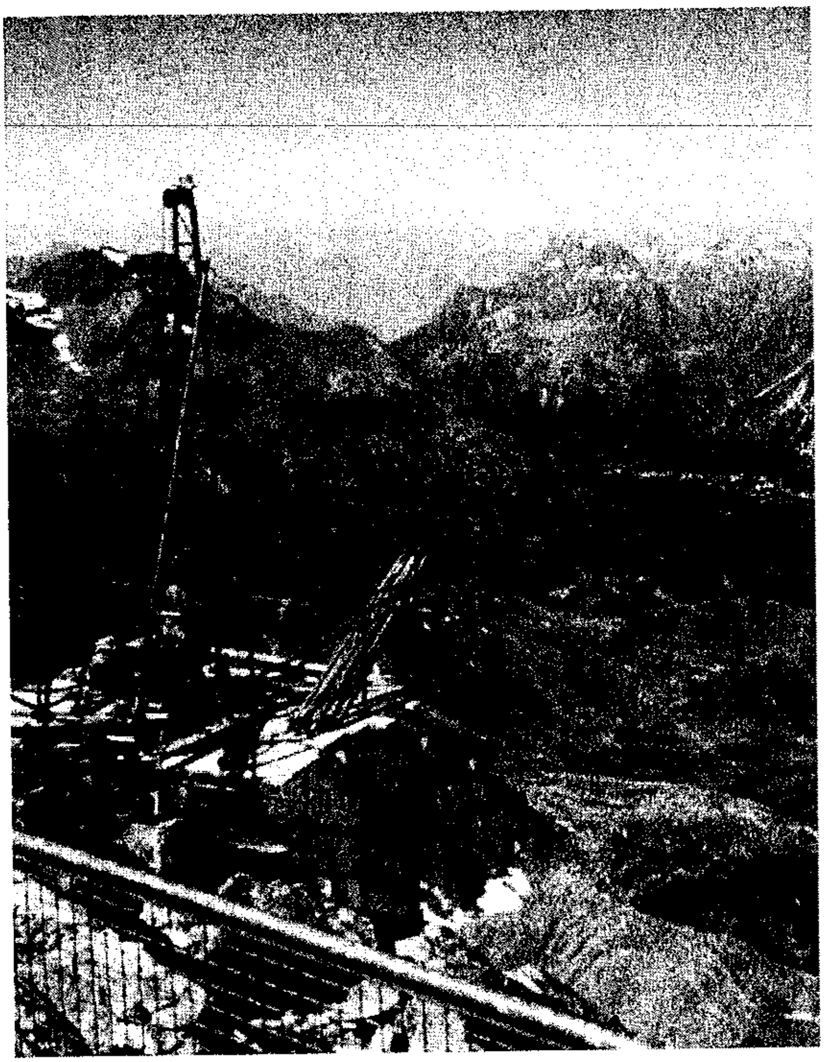

Figure 7. Drilling platform for the investigation of permafrost conditions at the Piz Corvatch cable station, 3303 m above sea level, Swiss Alps, in preparation for a new cable car terminal. Photo taken from Frozen Ground, Number 21, December 1997 (The News Bulletin of the International Permafrost Association).

THERMOKARST

The terrain response to permafrost thaw may be extremely rapid and dramatic. The processes and terrain instability associated with the thaw of permafrost are termed <u>thermokarst</u>.

Figure # 8 illustrates how ground subsidence and active-layer thickening may occur. Disturbance may be either natural, as in climate warming, or induced, such as by the modification of surface vegetation by man, fire, or other means. It is assumed that the permafrost is ice-rich. By this, we mean that the permafrost has a water content when thawed that is greater than saturation. The volume above saturation is termed 'excess ice'. Here, the figure assumes that the permafrost consists of 50% ice-saturated soil and 50% ice. If the active layer were originally 50 cm thick, a lowering of the base of the active layer by a further 50 cm would release 25 cm of saturated soil and 25 cm of excess ice. The subsequent melt of the excess ice leads to 25 cm of surface subsidence.

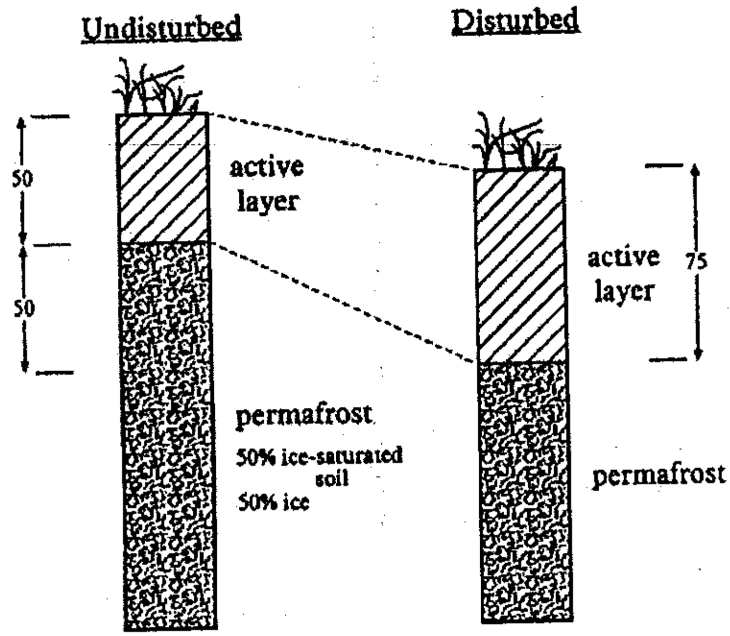

Figure 8. Schematic illustration of ground subsidence with active-layer deepening. In this case, the ice-rich permafrost has an excess ice content of 50%. If the active layer is originally 50 cm thick, then by lowering the base of the active layer a further 50 cm, 25 cm of saturated soil and 25 cm of excess ice are melted. Melting of the excess ice leads to 25 cm of surface subsidence. After Mackay, 1970; in Haeberli and Burn, 2002, figure 9.11.

A real example is provided by the ice-rich glaciolacustrine sediments that were illustrated in figure # 5. Figure # 9 plots the ground subsidence that occurred between 1994 and1999 at an experimental plot near Mayo, Yukon Territory. The original spruce forest cover was removed. Each year, the ground surface and the bottom of the active layer (frost line) were measured with reference to a nearby benchmark anchored in undisturbed permafrost. The ratio of subsidence to active layer deepening has been about 0.6:1. During the five years of observation, the surface subsided by 35 cm and the active layer thickened from 33 to 90 cm.

These data indicate the original active-layer thickness was 33 cm. If there were a small thickening of the active layer, say 25%, that occurred due to climate change, this would have caused ground subsidence of approximately 5 cm. While this may appear small, if the top of permafrost is characterized by high ice amounts, such deepening is sufficient to damage structures and buildings, and to initiate extensive landslides and other types of slope instability.

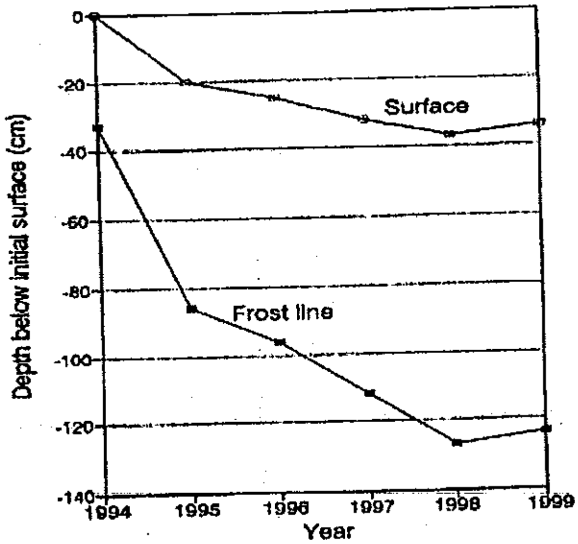

Figure 9. Graph showing ground subsidence (1994-1999) at an experimental plot in ice-rich glaciolacustrine sediments near Mayo, Yukon Territory, following surface disturbance. The absolute change in ground surface elevation and the top of frozen ground in late July-early August each year is indicated. From Haeberli and Burn, 2002, figure 9.12.

PAST PERMAFROST

Let me now talk about past permafrost. Although there is a broad relationship between the distribution of permafrost and present-day climate, it is equally obvious that, in many areas, the extent and thickness of permafrost bears little relationship to present-day climate.

The answer to this apparent paradox lies in the fact that much permafrost formed during the cold period of the Pleistocene. At these times, continental-scale ice sheets covered extensive areas of the northern hemisphere. These large ice sheets insulated the ground surface and may even have promoted thaw. However, in areas outside the limits of these ice sheets, permafrost formed and thickened progressively. For example, in parts of unglaciated Yukon and interior Alaska, a permafrost thickness in excess of 100 m characterizes the never-glaciated areas while a thickness of less than 60 m occurs within areas of glaciation. The remains of now-extinct Pleistocene-age fauna, such as woolly mammoths, also indicate the relict nature of much permafrost.

THE PLEISTOCENE PERIGLACIAL ZONE

During each glacial period, permafrost conditions extended southwards into the mid-latitudes of both North America and Eurasia. These cold ice-marginal conditions are termed 'periglacial'. In theory, a 'Pleistocene periglacial zone" existed between the southern limit of the continental ice sheets and the northern limit of trees. These concepts are illustrated schematically in Figure #10. The figure shows, on left, the theoretical limit of the periglacial zone as determined by climate and, in center, the Pleistocene periglacial zone, as displaced southwards and peripheral to the ice sheets. Thus, the present periglacial zone, on right, commonly incorporates a so-called 'relict periglacial zone' in which permafrost was once present.

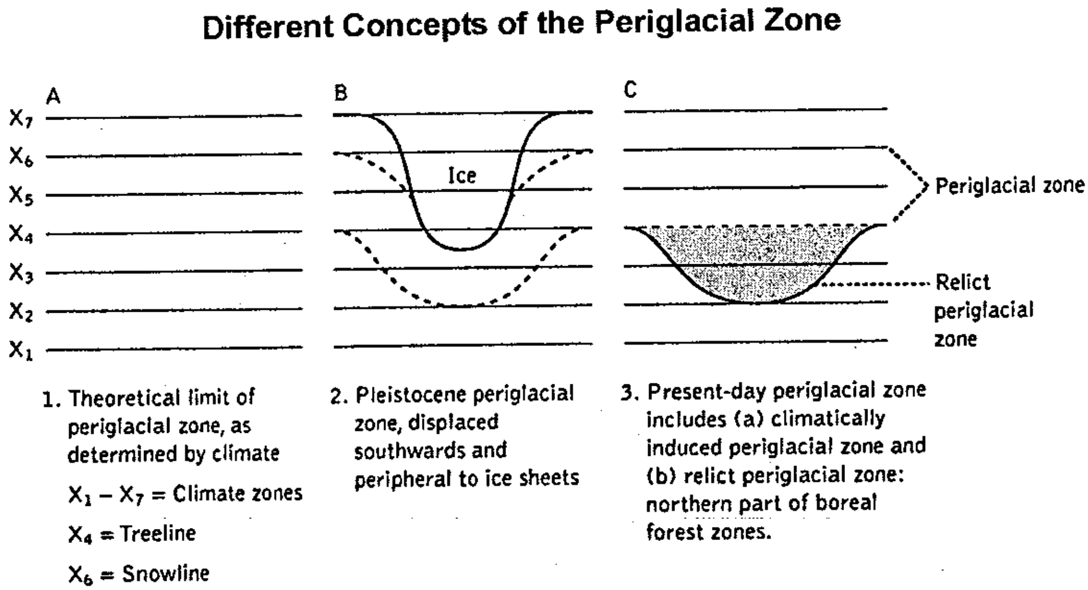

Figure 10. Schematic diagram illustrating different concepts of the mid- and high-latitude periglacial zone. From French, 1996, figure 1.2.

The Pleistocene periglacial zone of the mid-latitudes was also characterized by strong zonal winds, brought about by the constriction of the global wind belts, together with cold katabatic winds blowing directly off the ice sheets. They were especially effective in eroding and transporting sediment because the vegetation cover was sparse and often xeric.

The extent of the relict periglacial zone that existed in the United States at the height of the last glacial maximum, the Late-Wisconsinan, has been mapped by several workers. Figure #11 summarizes the traditionally-accepted extent of the Late-Pleistocene periglacial zone in the lower 48 states. It also indicates the different types of relict periglacial and permafrost-related phenomena that may now be found. You should note that southern New Jersey lies outside the relict periglacial zone, as depicted here. One objective of the 2003 GANJ field excursion is to demonstrate that southern New Jersey lay within the Late-Pleistocene 'periglacial zone', and that the region experienced intense frost, the formation of permafrost, and strong wind action, during at least two periods of the Late-Pleistocene.

PERMAFROST-RELATED LANDFORMS

In preparation for tomorrow's field excursion, let me briefly mention some of the interesting and unique landforms that are commonly associated with permafrost terrain today.

First, the lowering of temperature of ice-rich frozen soils leads to the thermal-contraction-cracking of the ground. Fissures, typically 1-4 m deep, become in-filled with varying combinations of ice and mineral soil, to form ice-, sand- or composite-wedges. The wedges form orthogonal or polygonal patterns at the ground surface, often with distinctive micro-relief. These are commonly termed tundra polygons. Their dimensions vary between 15 and 30 m. Undoubtedly, tundra polygons are the most widespread and ubiquitous of large-scale permafrost-related landforms.

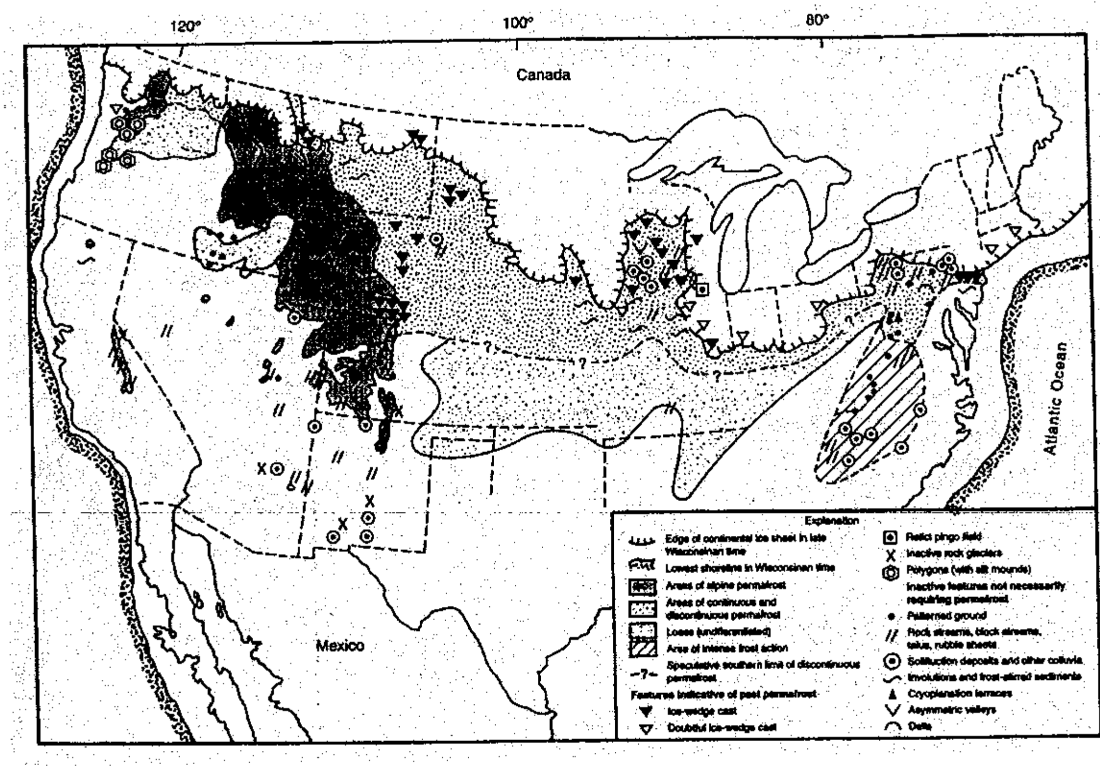

Figure 11. Tentative reconstruction of the maximum extent of Late-Wisconsinan periglacial conditions in the USA south of the ice sheet limit. From Péwé, 1983, figure 13.6.

Second, less widespread but equally distinct, are various frost mounds that result from the localized growth of ice bodies. The most well known of these are pingos, some of which may be 40 m high. These are formed either by ice injection under hydraulic pressure (open-system pingos) or through ice segregation and localized water intrusion under hydrostatic pressure (closed-system pingos). In all cases, pingos are discrete permafrost mounds that result from readily identifiable permafrost and hydrological conditions. Smaller frost mounds, such as seasonal frost mounds and palsas, are associated with more localized groundwater seepage or with peaty organic terrain.

Third, the thaw of permafrost (thermokarst) gives rise to distinct landforms and processes. For example, ground- ice slumps and thaw lakes and depressions result from the thaw-subsidence of icy permafrost. Furthermore, when the top of permafrost thaws, water-saturated sediments may initiate slope instability in the near surface. As thaw progresses and icy layers at depth are encountered, gravity-induced density readjustments may occur in the water-saturated sediments. These structures are termed 'thermokarst involutions'. The in-situ creep and deformation of coherent bedrock may also occur as the permafrost progressively warms. On slopes, mass movement, generally referred to as solifluction, occurs in the seasonally thawed layer.

Finally, the active layer itself is characterized by a number of frost-action processes. These result in the heave of the ground surface, the upfreezing and tilting of pebbles and stones, and particle-size sorting. Complex, frost-induced, circulatory mechanisms in this zone of seasonal freezing and thawing can lead to the formation of various sorted and non-sorted circles, and to small-scale patterned ground phenomena.

CONCLUSION

During the Late Pleistocene, southern New Jersey lay to the south of the large ice-sheets that covered much of the northern part of the North American continent. This ice-marginal 'periglacial' zone was characterized by intense frost, the formation of permafrost, the presence of strong winds, and sparse tundra-like vegetation. Today, there is evidence for the previous existence of such conditions in the Pine Barrens. This takes the form of ancient sand wedges, thermokarst structures, and numerous aeolian phenomena such as deflation hollows or depressions ('spungs'), wind-abraded rocks (ventifacts), and wind-polished boulders.

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Late Miocene to Holocene Geology of the New Jersey Coastal Plain

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ABSTRACT

The surface deposits and landforms of the New Jersey Coastal Plain record stepwise periods of fluvial incision and erosion separated by periods of fluvial and marginal-marine deposition. The major erosional periods correspond to sustained sea-level declines caused by glacial ice growth in Antarctica in the middle to late Miocene and in the northern hemisphere in the late Pliocene and early Pleistocene. The Beacon Hill and Bridgeton fluvial plains were deposited on the shelf surface exposed during late Miocene sea-level fall. The Pensauken fluvial plain was deposited during a period of high global sea level in the mid-Pliocene. During the repeated Laurentide glaciations of the middle and late Pleistocene, valley alluviation occurred during brief periods at peak glacial and peak interglacial times, separated by longer periods of gradual stream incision and valley widening.

INTRODUCTION

The New Jersey Coastal Plain (NJCP) is favorably situated to record the interplay of sea-level, climate change, and glaciation over the past 10 million years (m.y.). It is on an old, thermally mature passive margin, so there is little tectonic movement other than the gradual crustal adjustment to inland erosion and offshore deposition, which has probably produced no more than 60 feet of vertical change over the past 10 m.y. in this area (Pazzaglia and Gardner, 1994), and the more-rapid crustal depression and subsequent rise as the Laurentide ice sheet has grown and melted away about 10 times within the past 2 m.y. It is bordered on the north by the terminal positions of three glaciers, which were accompanied by periglacial climates in the NJCP. And, most importantly, it borders the Atlantic Ocean, the level of which has varied, along with that of all the world's oceans, over a range of about 600 feet in the past 10 m.y. as glaciers have grown and melted.

The surficial geology of the NJCP was first studied by Cook (1880), who recognized the general distribution of upland gravels and valley deposits, and who suspected the influence of both glacial and marine processes in their deposition. The first systematic mapping was by R. D. Salisbury and G. N. Knapp between 1891-1903. This mapping was partially published in Bascom et al. (1909a, b) and Johnson (1950) and was described narratively in Salisbury (1898) and, more comprehensively, in Salisbury and Knapp (1917). This work formally defined the Bridgeton, Pensauken, and Cape May formations, although their age and the relative roles of fluvial, glacial, and marine processes in their deposition remained uncertain. These roles were considerably clarified over the next 70 years. MacClintock and Richards (1936) and MacClintock (1943) established the marine origin of the Cape May Formation, and demonstrated that it was laid down during a Pleistocene interglacial highstand. Bowman (1966), Owens and Minard (1979), and Martino (1981) studied the paleoflow, sedimentology, mineralogy, weathering characteristics, and regional stratigraphic correlations of the Pensauken and Bridgeton formations and demonstrated their preglacial fluvial origin. Owens and Minard (1975, 1979) mapped the surficial geology in a part of the Coastal Plain southeast of Trenton, and reinterpreted the Pleistocene stratigraphy of the Delaware valley below Trenton. A cooperative effort of the New Jersey Geological Survey and the U. S. Geological Survey between 1982-1992 to produce a new geologic map of the state resulted in a 1:100,000 scale map of the surficial geology of the NJCP (Newell et al., 2000). A continuing effort by the New Jersey Geological Survey to map the state at 1:24,000 has produced detailed surficial geologic maps for the northern Coastal Plain (see publication listing and several viewable maps at www.njgeology.org) and the Camden area. The 1:100,000-scale map, with 1:24,000 updates, is being prepared for release on CD as a N. J. Geological Survey digital data set (figs. 1 and 2).

This paper will describe the lithology, sedimentary and weathering characteristics, evidence for age, and landscape setting for the principal surficial formations of the NJCP, and the sequence of erosional and depositional events they record. It will then tie these events to the global record of sea level and regional records of climate, river drainage, and glaciation.

SURFICIAL DEPOSITS

Beacon Hill Gravel

The Beacon Hill Gravel is a deeply weathered sand and gravel, composed of quartz, quartzite, and chert pebbles, that caps the highest hills in the NJCP. It occurs in small, widely separated erosional remnants, principally in the Mount Pleasant Hills in northern Monmouth County, the Clarksburg Hills in western Monmouth County, and the Woodmansie upland in the northern Pine Barrens in Ocean and Burlington counties. A quartzite-chert lag of pebbles and cobbles on summit flats on Tenmile Mountain and Rocky Hill in the Piedmont northeast of Princeton is also likely a residue from the Beacon Hill, and marks its northernmost preserved extent. Together, these erosional remnants define a surface for the base of the Beacon Hill that descends from 320 feet in the northern NJCP to about 190 feet in the Woodmansie area. Preserved remnants of the Beacon Hill are nowhere more than 30 feet thick, and the absence of the gravel lag on Rocky Hill above 350 feet in elevation also indicates a maximum Beacon Hill thickness of about 30 feet there. Thus, it is reasonable to reconstruct the Beacon Hill as a thin fluvial sand and gravel sheet on a generally flat, south-sloping plain.

Except in the Mount Pleasant Hills and in the Piedmont, the Beacon Hill rests on the Cohansey Formation. In most places the contact is sharp and channeled but in a few outcrops the contact is more transitional, leading some workers to consider the Beacon Hill as a fluvial facies of the Cohansey (Salisbury and Knapp, 1917; Newell et al., 2000). The Cohansey is a marginal-marine sand no older than latest middle Miocene, based on pollen (Owens et al., 1988) and a strontium stable-isotope ratio age of 12 Ma (million years ago) on the youngest

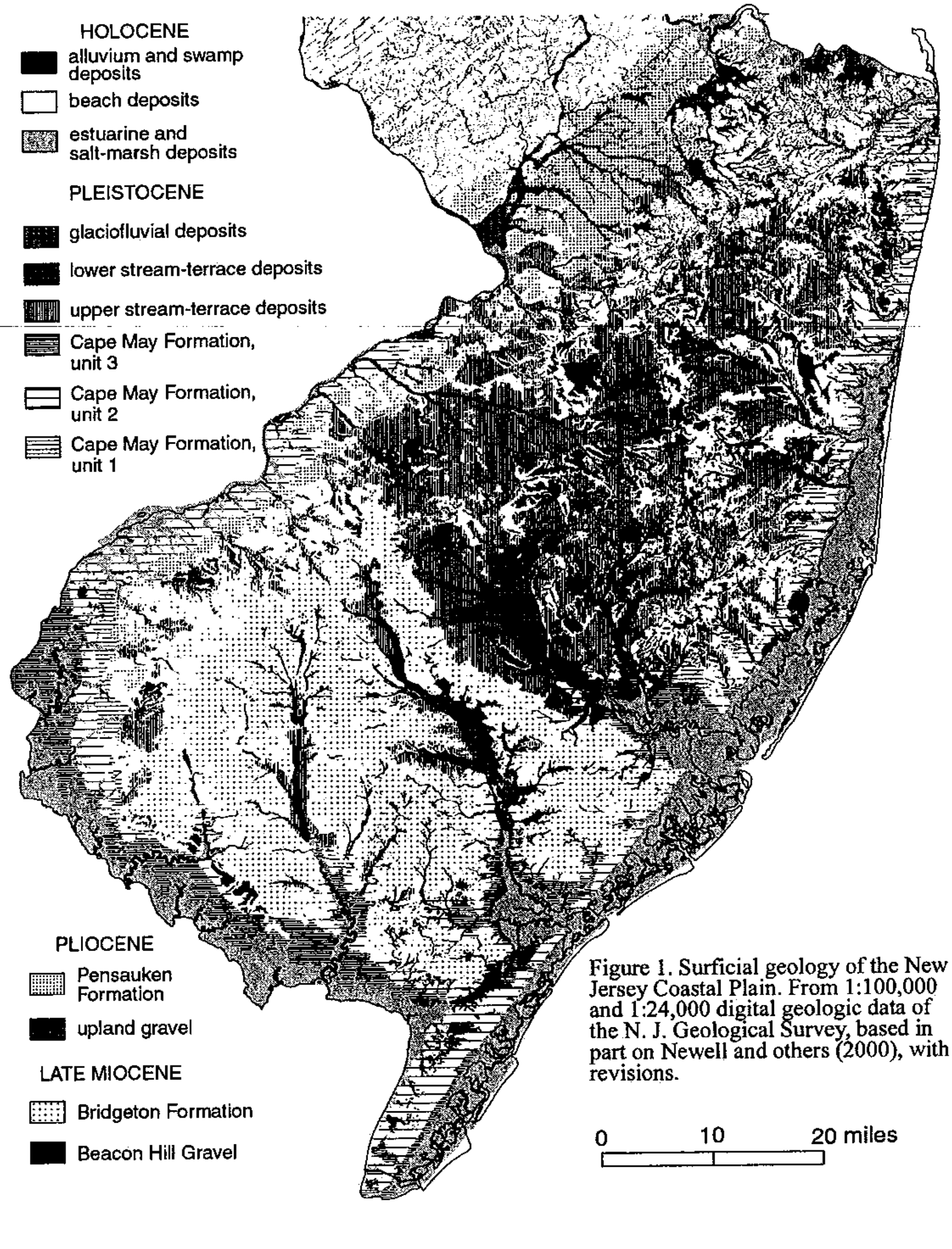

Figure 1. Surficial geology of the New Jersey Coastal Plain. From 1:100,000 and 1:24,000 digital geologic data of the N. J. Geological Survey, based in part on Newell and others (2000), with revisions.

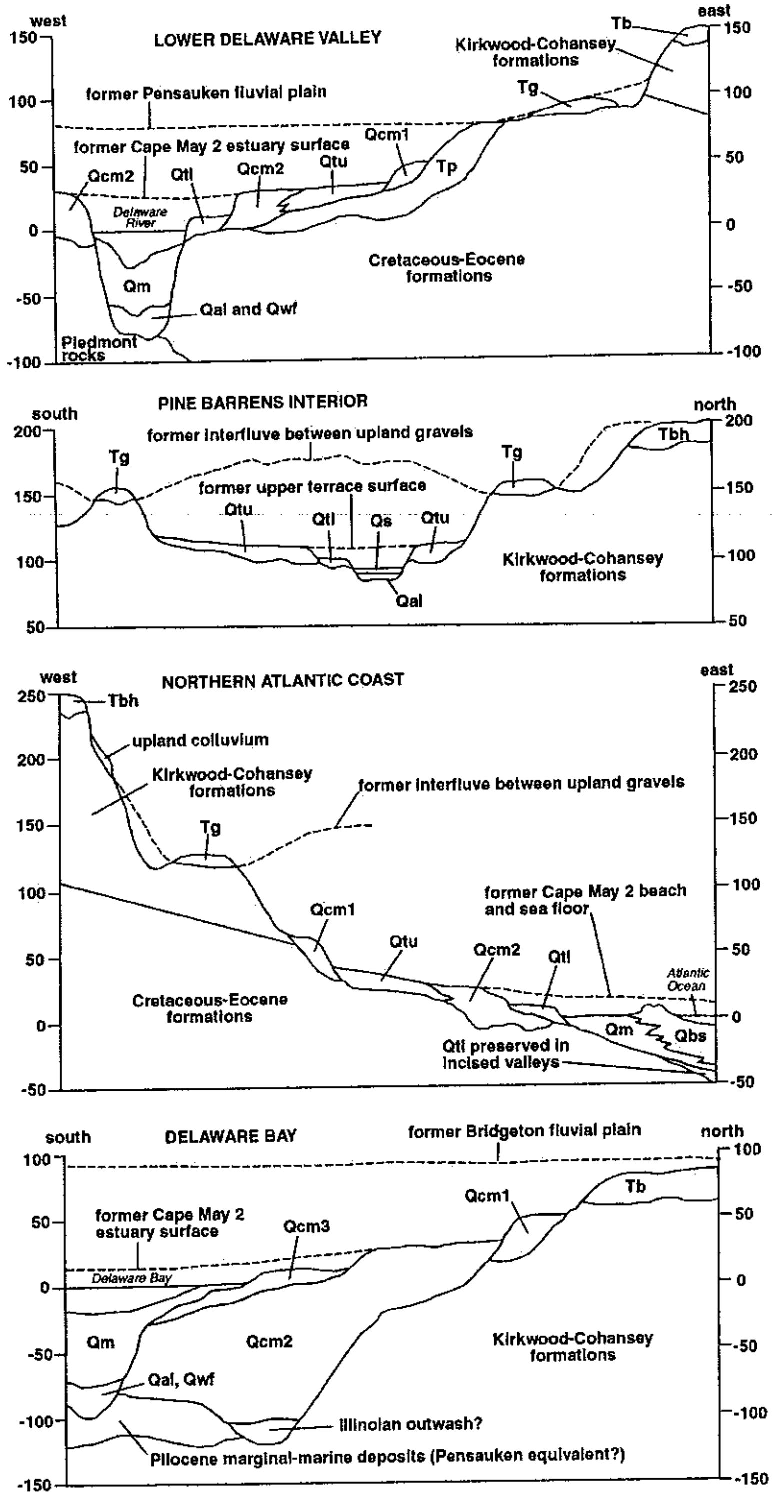

Figure 2. Schematic cross sections of various Coastal Plain terrains. Elevations in feet; great vertical exaggeration. Unit symbols are: Qm=salt-marsh and estuarine deposits, Qbs=beach sand, Qtl=lower terrace deposits, Qtu=upper terrace deposits, Qcm=Cape May Formation (units 1, 2, 3 indicated by suffix), Tg=upland gravels, Tp=Pensauken Formation, Tb=Bridgeton Formation, Tbh=Beacon Hill Formation. Dashed lines show former topographic or bathymetric surfaces.

member of the underlying Kirkwood Formation (Sugarman et al., 1993). The locally transitional contact of the Cohansey and Beacon Hill, and the flat, gently sloping surface on which the Beacon Hill rests, suggests that the Beacon Hill fluvial plain was deposited on the shelf surface exposed as sea level fell after deposition of the Cohansey. Indeed, the gradient of the Beacon Hill plain (0.0008) is similar to that of restored Miocene shelf surfaces (0.0007) on the New Jersey offshore margin (Steckler et al., 1999).

The Beacon Hill is chiefly structureless pebble gravel with a reddish yellow to yellowish red clayey sand to sandy clay matrix. The clay content and red color is largely from intense weathering of feldspar and iron-bearing mafic minerals, and the lack of bedding structure is the result of extended periods of biogenic, pedogenic, and cryogenic churning. The gravel is principally quartz, quartzite, and chert. Many of the cherts are weathered to a white or yellow chalky residue, and the quartz and quartzite are pitted and partially weathered along fractures and grain boundaries. There are also traces of gray and red sandstone and mudstone from Newark Basin and Appalachian rocks. These clasts are fully saprolitized. Where preserved, soils show well-developed microstructural features and a gibbsite-kaolinite clay mineralogy consistent with a long period of soil development in a warm, humid climate (Trela, 1984).

In a few outcrops, tabular-planar fluvial cross bedding is preserved. Current directions measured on these cross-beds indicate southerly paleoflow (Stanford et al., 2002) (fig. 3a), consistent with the southerly slope of the fluvial plain. Tidal-current flow direction and grain-size trends in the Cohansey Formation also indicate southerly transport of sediment in the northern NJCP (Carter, 1972), further supporting the possibility that the Beacon Hill is a regressive fluvial facies of the Cohansey, fed by the same south-flowing river system. Wind gaps in northern New Jersey, through the New Jersey Highlands and Kittatinny Mountain, are roughly on grade with the projected slope and flow trajectory of the Beacon Hill plain (fig. 3a), if some steepening of river grade is allowed inland of the Coastal Plain, as would be expected where channels are eroding into bedrock rather than depositing a plain on an exposed shelf surface. The topographic setting of the wind gaps and surrounding uplands suggests a landscape significantly flatter than the modern landscape of northern New Jersey, with ridges on resistant rock rising gently 200-400 feet above the gaps and intervening lowlands on softer rock.

The Beacon Hill is approximately dated to 10-9 Ma if one assumes that it was deposited shortly after the Cohansey, as the field evidence suggests. The Beacon Hill is much older than the Pensauken Formation, as indicated by the deep and extensive erosion separating them. The Beacon Hill plain was incised to a depth of as much as 300 feet, and almost completely eroded away, before deposition of the Pensauken Formation in the Pliocene, probably between 4 and 2 Ma. The soil and weathering properties of the Beacon Hill, and pollen and fossil plant material from the Cohansey Formation (Greller and Rachele, 1983) and from the Bryn Mawr and Brandywine gravels (McCartan et al., 1990; Pazzaglia et al., 1997), which are correlates of the Beacon Hill in the Maryland and Pennsylvania Piedmont, indicate a warm, temperate climate.

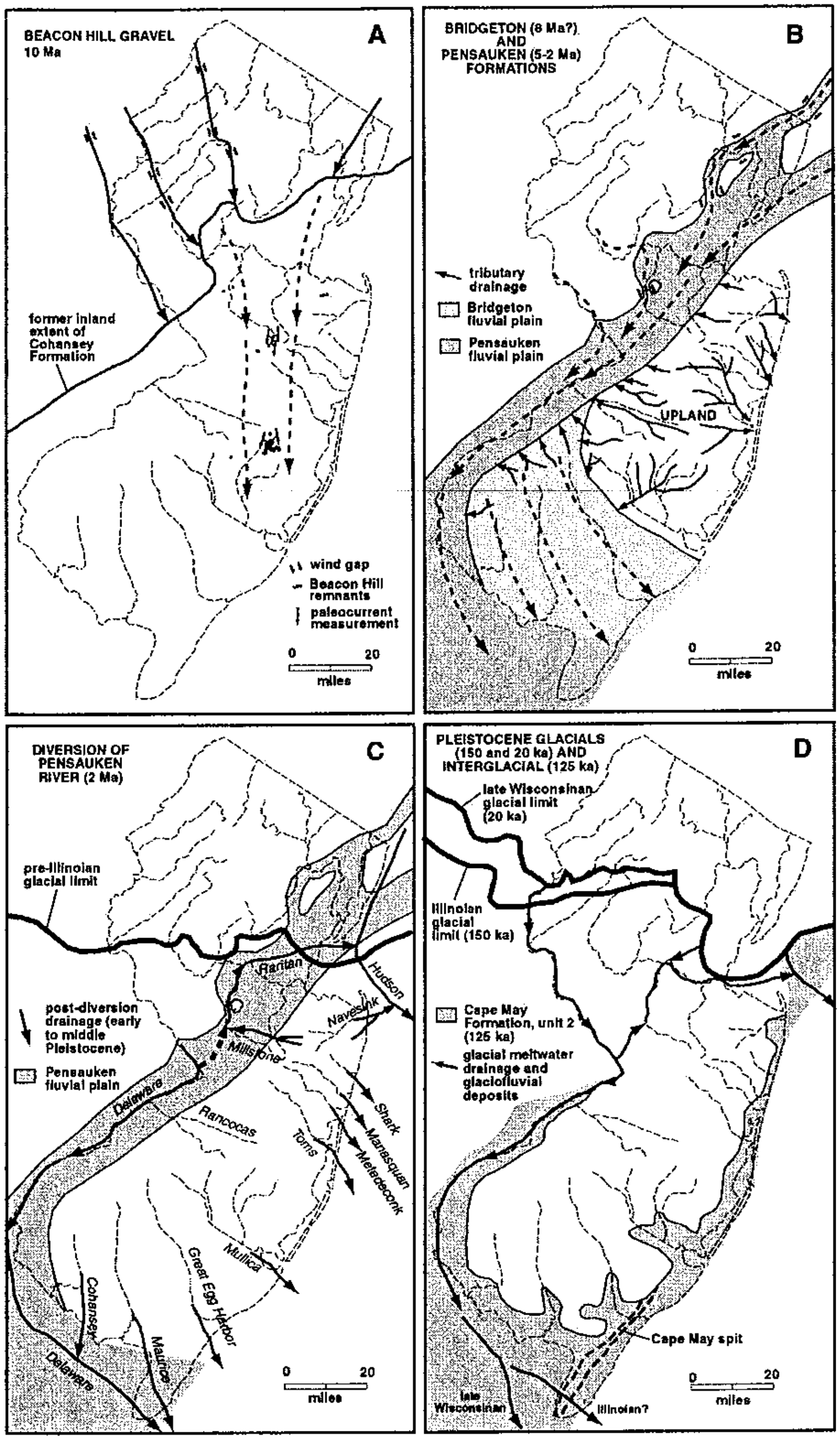

Figure 3. Events shaping the landforms and surficial deposits of the Coastal Plain. A) Beacon Hill fluvial plain is deposited on exposed shelf surface after Cohansey regression. B) Sequential deposition of Bridgeton and Pensauken fluvial plains. Tributary drainage marked today by upland gravels. C) Development of modern valleys after diversion of the Pensauken river during pre-Illinoian glaciation. D) Glacial meltwater routes during the Illinoian and late Wisconsinan glaciations and deposition of Cape May during the Sangamonian interglacial.

Bridgeton Formation

After the Beacon Hill plain was deposited, the continued fall of sea level allowed fluvial incision into the plain. This incision was accompanied by a reorientation of the southerly flow in Beacon Hill time to a more southwesterly flow in the northern Coastal Plain, sweeping around to a southeasterly flow across the southern Coastal Plain to the Atlantic (fig. 3b) (Owens and Minard, 1979; Martino, 1981). The cause of this reorientation is speculative. Owens and Minard (1979) suggest a tectonic control by a fault system running from the Hudson valley across central New Jersey to the Delaware valley, but no faults or folds have been detected in the Cretaceous formations in this area to support this hypothesis. The reorientation may have been a response to subsidence in the Salisbury embayment, a marine depositional center offshore of what is now the Chesapeake Bay-Delmarva Peninsula area. This embayment has a thicker accumulation of Miocene through Pleistocene marine sediment than the New Jersey area (Poag and Sevon, 1989), indicating greater subsidence there during this period. This subsidence may have imparted a southwesterly tilt to the NJCP that shifted the river flow.

The reoriented drainage deposited a fluvial plain greater than 50 km wide across the southern Coastal Plain, south of the modern location of the Mullica River valley. This plain is composed of reddish yellow tabular-planar to trough cross-bedded clayey sand, sand, and pebble gravel, with some cobbles, known as the Bridgeton Formation. The base of the Bridgeton descends from an altitude of 150 feet in the Berlin area of eastern Camden County to about 40 feet along the Atlantic coast. It is generally less than 25 feet thick but locally is as much as 40 feet thick. North of Berlin the Bridgeton is absent. The northward extension of the Bridgeton plain was presumably much narrower than that to the south of Berlin and was eroded away during the period of incision prior to deposition of the younger Pensauken Formation. Owens and Minard (1979) extend the Bridgeton north of Berlin to the Perth Amboy area, reassigning deposits formerly included in the Pensauken Formation by Salisbury and Knapp (1917). This reassignment was followed by Martino (1981) and Stanford (1993) and in part by Newell et al. (2000). However, the elevation of the top of these reassigned deposits is 150-200 feet below the projected base of the Bridgeton plain. This discrepancy is possible only if folding or faulting had down-dropped the northern portion of the Bridgeton plain greater than 200 feet relative to the plain south of Berlin. The underlying Coastal Plain strata show no such offset, so the reassignment of the northerly deposits to the Bridgeton is untenable. The elevation and paleoflow direction of these deposits clearly place them on grade with those of the Pensauken type area, and thus the original mapping by Salisbury and Knapp (1917) is retained.

Like the Beacon Hill, the Bridgeton is intensely weathered and supports deep, mature soils (Tedrow, 1986). Again, the clay content of the matrix is largely from decomposition of feldspar minerals in the sand. The gravel fraction includes mostly quartz and quartzite, with some chert, and traces of igneous and metamorphic rock from the Piedmont, and sandstone and mudstone from the Appalachians and Newark Basin. These non-quartz clasts are commonly saprolitized or have thick weathered rinds. The gravel composition is similar to that of the Beacon Hill, although there is less chert and more metamorphic and igneous rock. These differences may be due to the greater availability of fresh rock in Bridgeton time owing to stream incision in the Piedmont, or to a greater contribution from streams from the west draining across the Pennsylvania Piedmont, or to the less-weathered and eroded state of the Bridgeton, which has preserved more of the igneous and metamorphic rocks.

The upper 5 to 10 feet of the formation are usually structureless or show cryoturbation features such as frost wedges and involutions. Below this bioturbated and cryoturbated zone, the Bridgeton is chiefly a planar-tabular to trough cross-bedded sand and pebble gravel with some local channel lags of pebble-to-cobble gravel. Abundant current measurements on the cross beds clearly show the southeasterly paleoflow of the river system (Owens and Minard, 1979; Martino, 1981). Despite the intense weathering, the Bridgeton is still largely intact on the southern Coastal Plain away from the Delaware valley, and the topography of the fluvial plain readily can be reconstructed.

The age of the Bridgeton is bracketed by its relationship to the Beacon Hill-Cohansey and Pensauken formations. It unconformably overlies the Cohansey and older Coastal Plain formations and is erosionally inset into the Beacon Hill. It, in turn, was incised as much as 150-200 feet, and extensively eroded in the Delaware valley and Delaware Bay area, before deposition of the Pensauken. Although the Bridgeton is inset into the Beacon Hill, it is not deeply inset and the two deposits may not be widely separated in time. The base of the Bridgeton grades to about 40 feet along the Atlantic coast in the Atlantic City area; the base of the Beacon Hill plain projects to about 75 feet in the same area. If the base of the Bridgeton plain is projected north and northeastward up the Delaware valley from Berlin, it is at an elevation of about 320 feet in the Clarksburg Hills, nearly equivalent to the base of the Beacon Hill there. Thus, the Bridgeton can be viewed as a topographically lower phase of the Beacon Hill, related to the same long-term sea-level decline in the late Miocene after Cohansey deposition. An age estimate of 7-9 Ma may be appropriate.

Climate during Bridgeton time was a continuation of the warm-temperate climate prevailing in Cohansey-Beacon Hill time. The only direct evidence of climate are plant fossils collected from the Bridgeton near Bridgeton (Hollick, 1892). These include oak, blackgum, holly, hickory, magnolia, beech, and sweetgum, trees characteristic of the southern Atlantic Coastal Plain today.

Upland Gravels

With the southwestward drainage shift recorded by the Bridgeton, the northeastern NJCP began to develop as an upland that formed a divide between the Hudson-Delaware valleys and the Atlantic Ocean (fig. 3b). Local streams drained out radially from this upland to the Bridgeton plain to the south and west and to the Atlantic to the east. The extent of this upland to the northeast has been erased by subsequent glacial, fluvial, and marine erosion. It likely extended northeastward to include what is now Long Island, as indicated by the evidence for a southwesterly flowing river in the Long Island Sound area, on the inland side of the upland, in the Pliocene (see section on Pensauken Formation below).

As the local streams incised into the upland, they reworked the Beacon Hill Gravel capping the highest plateaus in the area and deposited sand and gravel in channels and floodplains. With continued incision through the Pliocene and early Pleistocene, these channel gravels were more resistant than the bordering sand of the Cohansey and Kirkwood formations, which underlie most of the upland and formed the valley sides. Streams thus cut down in the sand rather than in the channel gravels, eventually leaving the gravels as ridges or hills in a process known as *inversion*. These ridgetop deposits are collectively called *upland gravels*, and they mark the former courses of streams. Although many are too eroded and fragmented to be confidently fit into a drainage pattern, others can be topographically traced to where they grade to either the Pensauken or Bridgeton fluvial plains, or as they descend eastward to the Atlantic. These traces thus mark the location of the former tributary valleys (fig. 3b).

The upland gravels are typically almost entirely composed of quartz and quartzite, and local ironstone eroded from Coastal Plain formations, with only a trace of chert. This gravel composition reflects reworking of the Beacon Hill and, for those upland gravels tributary to the Pensauken in the lower Delaware valley, reworking of the Bridgeton. The deeply weathered chert, metamorphic, and sedimentary rocks in the Beacon Hill and Bridgeton did not survive erosion and transport while the unweathered or lightly weathered quartz and quartzite did survive. The sand and gravel are typically less than 10 feet thick and structureless owing to weathering, soil development, cryoturbation, and bioturbation. In places the sandy parts of the deposit are thinly layered and contain or rest on pebble lags, features that suggest deposition from sheetwash at the base of hillslopes. Elsewhere, the upland sand and gravel deposits are cross-bedded, indicating fluvial deposition. Newell et al. (2000) consider part of the upland gravels, and some Beacon Hill remnants, to be residuum derived from weathering and erosion of underlying Coastal Plain formations. This interpretation is untenable because these deposits contain gravel not present in the underlying formations, and, as noted above, have bedding structures indicating fluvial and colluvial transport.

In a few places, for example, in the Mount Pleasant Hills and Clarksburg Hills in Monmouth County, erosional remnants of the upland gravels are large enough to give an impression of the overall topography at the time of their deposition. The remnants suggest broad, flat-floored valleys with bordering pediments rising gradually 50-100 feet to Beacon Hill-capped mesa-like uplands. This valley form is similar to the broad valleys common in most of the NJCP today, and may be characteristic of valleys shaped by groundwater seepage from the base of adjacent uplands (Stanford et al., 2002).

Pensauken Formation

Continued decline of sea level after deposition of the Bridgeton Formation led to a renewed round of fluvial incision and drainage shift to the southwest. A new valley was eroded on the western and southern edges of the Bridgeton plain, and northeastward from the Camden area through Trenton and New Brunswick to the Jersey City-Manhattan area (fig. 3b). The thalweg of this valley is generally keyed into the contact of Cretaceous sediments and Piedmont or Newark Basin rocks. Southwest of the Princeton area, the deepest part of this valley extends below modern sea level (Stanford et al., 2002), indicating that it was cut during a period of lower-than-present sea level.

In the Newark Basin, the courses of rivers eroding this valley are marked by wind gaps. These include gaps through the Palisades Ridge at Sparkill, New York, paired gaps through the double Watchung ridges at Paterson and Millburn, and gaps through Third Watchung Mountain at Pine Brook and Livingston, all cut by a proto-Hudson river; a broad gap through the Palisades Ridge between Union City and Staten Island, cut by the trunk Pensauken river flowing from the present location of Long Island Sound; and a gap through Rocky Hill cut by a south-flowing proto-Raritan that was a tributary to the main Pensauken river (Stanford, 1993).

After the valley was cut, fluvial sand and gravel aggraded in the valley to form a plain as much as 15 miles wide. This deposit is the Pensauken Formation. It is as much as 140 feet thick in eastern Mercer and southern Middlesex counties, where it has not been deeply eroded, but is generally less than 40 feet thick elsewhere, where it was eroded during Pleistocene incision of the Delaware and Raritan rivers. The surface of this plain declines from 160 feet in the New Brunswick area to 80 feet in the Salem area. South of Salem, the plain broadened and spread southeastward, covering most of what is now the northern Delmarva Peninsula and probably extending across what is now Delaware Bay. Abundant paleocurrent data show southwesterly flow down the valley in New Jersey, sweeping to the south and southeast where the plain broadens on the Delmarva Peninsula (Owens and Minard, 1979; Martino, 1981; Stanford et al., 2002).

The Pensauken Formation is a tabular-planar to trough cross-bedded sand and pebble gravel, with minor cobble-gravel channel deposits at the base. Gravel is predominantly quartz and quartzite, with a little chert and a trace of igneous and metamorphic rock, sandstone, and mudstone, especially in the basal beds. Like the upland gravels, most of the gravel is reworked from the Beacon Hill and Bridgeton formations. As in the upland gravels, chert content is reduced from that in the Beacon Hill and Bridgeton because chert pebbles were weathered and did not survive reworking. The sand fraction is quartz with some feldspar and heavy minerals, and, in places, glauconite and mica (Bowman, 1966). Glauconite is most abundant on the southeastern side of the Pensauken valley, where tributaries draining from the inner Coastal Plain were eroding glauconite-bearing Cretaceous through Eocene formations. Glauconite is much less abundant in the Beacon Hill and Bridgeton formations because the glauconite-bearing formations were largely covered by the nonglauconitic Kirkwood and Cohansey formations when the Bridgeton and Beacon Hill were deposited.

Pollen from eight outcrop and borehole samples of a black clay bed within the Pensauken near Plainsboro (table 1) include the pre-Pleistocene exotic taxa *Engelhardia* and *Pterocarya* but lack additional exotic taxa characteristic of Miocene and older deposits in this region. This assemblage is typical of the Pliocene in the mid-Atlantic Coastal Plain (Groot, 1991; Pazzaglia et al., 1997). Also favoring a Pliocene age for the Pensauken is its relation to pre-Illinoian till, which overlies the Pensauken near Somerville in the Newark Basin. This till is weathered and eroded to a degree similar to the Pensauken. The pre-Illinoian till correlates westward to magnetically reversed drift in central Pennsylvania (Gardner et al., 1994), indicating a pre-788 ka (thousands of years ago) age, and pre-Illinoian lake clay in northern New Jersey contains pre-Pleistocene exotic pollen taxa similar to those in the Pensauken (Harmon, 1968; Stanford, 1997). These facts suggest that the pre-Illinoian drift in New Jersey correlates to the late Pliocene K Laurentide glaciation of Richmond and Fullerton, which is dated by ash and magnetic stratigraphy in the Missouri valley to 2.1 Ma (Richmond and Fullerton, 1986). Before deposition of the Pensauken, the extensive period of erosion following deposition of the Beacon Hill and Bridgeton formations indicates a significant time gap between those formations and the Pensauken.

Table 1. Pollen and radiocarbon data for the NJCP. Numerous radiocarbon dates younger than 8 ka from organic material in Holocene alluvial, swamp, and estuarine deposits are not listed. Unreferenced sites are from Stanford et al. (1992).

Site or Reference	Location latitude-longitude, depth, and material	Stratigraphic Position[1]	Radiocarbon Date[2], lab number, and $\delta^{13}C$ (‰)	Pollen (abundant taxa in boldface, other common taxa in normal font)
Above Buck Run bog (Stanford, 2000; Russell, 2000)	39°42'07", 74°30'05", 2.2 m, peat	base of Qs	8,050±40 GX-26535-AMS -27.2	**pine, oak**, cedar
Old Oxbow bog (Stanford, 2000)	39°40'30", 74°42'49", 2.2 m, peat	base of Qs	8,900±90 GX-26537, -27.9	
McDonalds Branch bog (Buell, 1970)	39°53'11", 74°30'18", 1.3 m, peat and wood	base of Qs	9,125±195	
Helmetta bog (Watts, 1979)	40°23', 74°26', 2.1 m, peat	base of Qs	9,640±60 QL-1082	**pine**, oak
90-1373	40°17'52", 74°41'47", 1.2 m, peat	base of Qal	10,430±160 I-16554	
Oswego River bog (Buell, 1970; Florer, 1972)	39°42'50", 74°31'34", 2.4 m, peat	base of Qs	10,485±240	**pine, oak**, hemlock, spruce
CO149 (Sirkin et al., 1970)	40°02'46", 74°40'23", 1.5 m, wood	base of Qal	10,770±300 W-1411	**pine**, spruce, hemlock
boring B2 Raritan estuary	40°30'02", 74°17'05", 29 m, organic silt	base of Qm	11,420±560 GX-21687, -26.6	**oak, pine, hemlock, birch**, alder, spruce
Szabo Pond (Watts, 1979)	40°24', 74°29', 2.1 m, organic silt	within Qs	11,950±100 QL-965	**spruce, pine**, birch
96-186	40°22'56", 74°13'25", 1.8 m, wood	base of Qal	12,700±170 GX-21686, -25.3	**spruce, alder**, pine, birch, traces of oak and willow, moderate grass and sedge in non-arboreal fraction
TW21 (Sirkin et al., 1970)	40°08'45", 74°50'42", 2.4 m, peat	base of Qst	13,200±400 W-1195	**pine, birch, spruce**, alder
SH29 (Sirkin et al., 1970)	40°23'41", 74°04'51", 1.2 m, peat	base of Qal	13,680±300 W-2119	**spruce, pine**, birch, alder; abundant grass and sedge in non-arboreal fraction
Minard (1969)	40°23'41", 74°04'51", depth not reported, peat	base of Qal	14,150±450 W-1457	
TE24 (Sirkin et al., 1970)	40°09'21", 74°39'22", depth not reported, wood	base of Qtl	26,800±1000 W-1193	**pine**, spruce, birch, alder, willow, blackgum, chestnut, oak
Marcus Hook, PA BH-1 (Jengo, 1999)	39°49'10", 75°24', 9.1 m, wood	within or reworked from Qtl	31,380+4530-2880	
core SO8 (Gaswirth, 1999)	40°26'26", 74°02'31", 8.2 m, wood	base of Qtl	31,740±1830 Beta 90133, -25.0	
SB-1 (Newell et al., 1995)	39°43'53", 74°13'56", 8.8 m, peaty clay-silt	base of Qtl	34,840±960 GX-16789-AMS	**alder, pine**, maple, birch, spruce, cedar, traces of chestnut, ash, blackgum; abundant sphagnum
96-308	40°11'10", 74°13'25", 1.8 m, organic silt	base of Qtl	35,570+3180-2270 GX-24257, -25.7	**oak, pine, spruce**, hickory, beech, blackgum
92-349	40°15'57", 74°27'34", 1.9 m, peat	within Qtu	>36,000 GX-19226, -23.8	**reworked Cretaceous pollen**, blackgum, oak, pine, cedar, hornbeam
SB-1 (Newell et al., 1995	39°43'53", 74°13'56", 18.3 m, organic clay-silt	within Qcm2	>39,000	

Site or Reference	Location latitude-longitude, depth, and material	Stratigraphic Position [1]	Radiocarbon Date [2], lab number, and $\delta^{13}C$ (‰)	Pollen (abundant taxa in boldface, other common taxa in normal font)
Artificial Island (Owens and Minard, 1979)	39°27'40", 75°32', 12 m, organic clay	within Qcm3 or Qcm2	>40,000 W-2296	
PG-10 (Newell et al., 1995)	40°00'15", 75°25'46", 11 m, organic clay-silt	Qcm2	>40,000	**hickory, hemlock, pine, oak, blackgum**, birch, beech, willow, ash, chestnut
CAN-3 (Newell et al., 1995)	39°28'37", 75°26'25", 2.4 m, organic clay-silt	Qcm3	>40,000	**oak, pine**, hemlock, hickory, birch, alder, blackgum, sweetgum, spruce, trace holly, beech, elm, fir, basswood, ash
Artificial Island (Owens and Minard, 1979)	39°27'40", 75°32', 14 m, wood	within Qcm3 or Qcm2	>42,000 W-2266	
96-179	40°19'02", 73°50'00", 1.8 m, organic clay	within Qcm2	>40,470 GX-21685, -20.6	**oak, pine, hickory, beech**, birch, sweetgum, cedar, spruce, abundant grass and some Chenopodiaceae/Amaranthaceae in non-arboreal fraction
PG-10 (Newell et al., 1995)	40°00'15", 75°25'46", 17 m, organic clay-silt	within Tp (beneath Qcm2)	>40,000	**alder, pine, oak, hickory, hemlock, blackgum**, maple, ash, elm, sweetgum, plus the pre-Pleistocene exotics *Momipites, Gordonia?, Plicapolis*
91-1224	40°20'59", 74°33'38", 4.6 m, organic clay	within Tp		**pine, spruce**, birch, hemlock, larch, oak, fir
boring HT2	40°21'04", 74°32'50", 6.4-9.7 m, organic clay	within Tp		6.4-6.7 m: **pine, oak**, hickory, hemlock, birch, *Engelhardia, Pterocarya*, trace *Micrhystridium* (dinoflaggelate) 6.7-7 m: **pine, hickory**, *Pterocarya*, oak, spruce, birch, alder, linden, hemlock, hornbeam, trace *Engelhardia* 7-7.3 m: **pine**, hickory, *Pterocarya*, hemlock, oak 8.2-8.5 m: **pine, hickory**, spruce, hemlock, oak, linden, fir, alder, birch 8.8-9.1 m: **pine**, hemlock, fir, birch, hickory, beech, trace of acritarchs 9.4-9.7 m: **pine**, birch, *Pterocarya, Engelhardia*, oak, cypress
boring HT3	40°20'20", 74°30'27", 11.3-12.2 m, wood	within Tp	>50,000 GX-17541, -28.9	**pine**, hickory, oak, hemlock, birch, juniper, abundant grass and some Chenopodiaceae in non-arboreal fraction, trace *Micrhystridium*

[1] Stratigraphic units are: Qal=modern floodplain deposits, Qs=modern swamp deposits, Qm=modern estuarine deposits, Qst=postglacial stream-terrace deposits, Qtl=lower terrace deposits, Qtu=upper terrace deposits, Qcm2, 3= Cape May Formation, unit 2 or 3, Tp=Pensauken Formation.
[2] All dates in ^{14}C yr B.P. Where not cited, laboratory number, error, and $\delta^{13}C$ value are not reported in the original source.

Thus, an age of 4-2 Ma is reasonable for deposition of the Pensauken plain. This interval corresponds to a period of high global sea level around 3.5 Ma, which is recorded in the middle Atlantic Coastal Plain by the Beaverdam Formation of the southern Delmarva Peninsula and the Yorktown and Duplin formations of Virginia and the Carolinas (Dowsett and Cronin, 1990; Krantz, 1991). The aggradation of the Pensauken plain following erosion of the Pensauken valley is the response of the Pensauken river system to this rising sea level. Indeed, the clay bed at Plainsboro, at an elevation of about 70 feet, contains a trace of brackish-water acritarchs, suggesting estuarine influence (Stanford et al., 2002). Pollen (table 1) and plant fossils (Berry and Hawkins, 1935) from the Pensauken indicate warm-temperate climate during Pensauken aggradation.

Diversion of the Pensauken River

The pre-Illinoian glacier advanced to a terminal position across north-central New Jersey (fig. 3c). East of the Watchungs the pre-Illinoian deposits are overprinted and eroded by late Wisconsinan glacial deposits, so their eastward extent is uncertain. However, it is clear that the pre-Illinoian glacier extended across the Pensauken valley in the New York City area. After retreat of this glacier, the trunk Pensauken river, which had flowed from southern New England via what is now the Long Island Sound lowland, and the proto-Hudson river, which formerly flowed across northeastern New Jersey as a tributary to the trunk Pensauken river, were rerouted to the Atlantic across the former upland to the southeast of the Pensauken valley. This rerouting may have been the result of glacial erosion of the upland, or may have happened by stream capture from local drainage eroding headward through the upland. Whatever the mechanism of diversion, the Pensauken-Hudson drainage following the pre-Illinoian glaciation was southeastward to the Atlantic from the New York City area. The Pensauken plain between the New York City and Trenton areas was abandoned, and a new local drainage system leading eastward to the Hudson was established, including the lower Raritan and Millstone rivers (fig. 3c). At Trenton, the Delaware turned southwest and drained down the former Pensauken valley, to which it had previously been a tributary.

Following these drainage rearrangements, the streams began to incise valleys into the former Pensauken plain, to a depth of as much as 80 feet below the former surface of the plain. This incision created the valleys in which modern streams flow, and led to topographic inversion of stream gravels in the Coastal Plain that were formerly tributary to the Pensauken. These gravels are a lower phase of the upland gravels.

Upper Terrace Deposits

Present-day valleys in the NJCP are dominated in most places by a broad upper terrace with a surface ranging from 20 to as much as 50 feet above the modern floodplain. These deposits rest on straths and pediments cut into underlying Coastal Plain formations and are generally less than 20 feet thick. They are chiefly quartz sand, with some glauconite in the inner Coastal Plain, and quartz-quartzite pebble gravel, with some ironstone and little to no chert. The sands and gravels are reworked from older surficial deposits and Coastal Plain formations, again with the loss of weathered components upon reworking. The sand is generally massive from cryoturbation and bioturbation. Where bedding is preserved it is typically planar-laminated to low-angle ripple cross-bedded, with pebbly channel lags. These structures suggest deposition in broad, shallow channels or from sheet flows. At the foot of some steep slopes (generally >8°) bordering the terraces, deposits of sand to silty sand colluvium, generally less than 10 feet thick, grade to the terrace surface. These deposits are massive to weakly layered, and rest on and contain pebble lags. They are of much smaller extent than the terrace deposits.

Along the Atlantic coast, and Delaware Bay and estuary, the upper terraces grade to, or are onlapped by, the Cape May 2 marine-estuarine deposit (fig. 4). Upper-terrace surfaces commonly have thermokarst basins and frost involutions and, in places, polygonal patterned ground, indicating that they were in place and stable during the last glacial maximum at 20 ka. A radiocarbon date of >36 ka (table 1) also shows that these terraces predate the last glacial maximum. These observations, and the relationship to the Cape May 2, indicate that the upper terraces are of Sangamonian (125 ka, oxygen-isotope stage 5) or older age. By analogy to the well-dated lower terrace deposits, the upper terrace deposits were probably laid down primarily during the peak Illinoian glaciation (150 ka), when the Illinoian glacier advanced into northern New Jersey and permafrost in the NJCP enhanced surface runoff and alluvial deposition. However, some of the upper terrace deposits, particularly those just inland from the Cape May 1 and 2 marine terraces, may be fluvial-estuarine backfills deposited in valleys at the head of estuaries during the Sangamonian and earlier sea-level highstands.

Newell et al. (2000) split the upper terraces into 2 or 3 subunits based on elevation as shown on 1:24,000 topographic maps. However, analysis of these same areas on 1:12,000 stereo aerial photographs and by field observation indicates that there are no recognizable scarps or treads within the upper terraces that would permit identification and mapping of multiple terraces. Rather, the elevation variation of the upper-terrace surface has a form more consistent with gradual degradation of a single surface during subsequent stream incision. The large size of some of the terraces (for example, those in the lower Rancocas River valley) indicates that parts of them are likely of pre-Illinoian age, but these are not topographically distinct from those of Illinoian-Sangamonian age.

Cape May Formation

A nearly continuous deposit of marginal marine-estuarine origin rings the coastal areas of New Jersey from Rumson southward along the Atlantic coast and Delaware Bay shore, forming the Cape May peninsula and extending up the Delaware estuary to the vicinity of Burlington (fig. 3d). These deposits are collectively known as the Cape May Formation (Salisbury and Knapp, 1917). They include isolated erosional remnants of shorefront terraces up to 70 feet above sea level, designated the Cape May 1 (Cape May 3 of Newell et al., 2000), a continuous, well-developed shorefront terrace with a surface between 10 and 35 feet above sea level, designated the Cape May 2, and, along the Delaware Bay shore and locally along the Atlantic coast, a lower shorefront terrace with a surface between 5 and 15 feet above sea level, designated the Cape May 3 (Cape May 1 of Newell et al., 2000). Along the Delaware Bay shore, the Cape May 3 is separated from the Cape May 2 by a scarp known as the Cedarville scarp.

The Cape May deposits are typically laminated to trough cross-bedded quartz sand and quartz-quartzite pebble gravel that are generally less than 40 feet thick. The sand includes some glauconite in the Delaware valley and northern Atlantic coast. The bedding structures are indicative of beach, tidal channel, tidal delta, and nearshore marine environments. Beneath the Cape May peninsula, and parts of the Delaware Bay and adjacent bayshore, the deposits thicken to as much as 150 feet, and include an upper sandy beach-spit facies and a lower silty-clayey estuarine-bay facies (Gill, 1962: Lacovara, 1997). The bay-estuarine facies in part fills a paleovalley cut by the Delaware during the Illinoian sea-level lowstand (Knebel and Circe, 1988) (fig. 2). Silt and clay are present elsewhere in the Cape May, but are generally minor constituents.

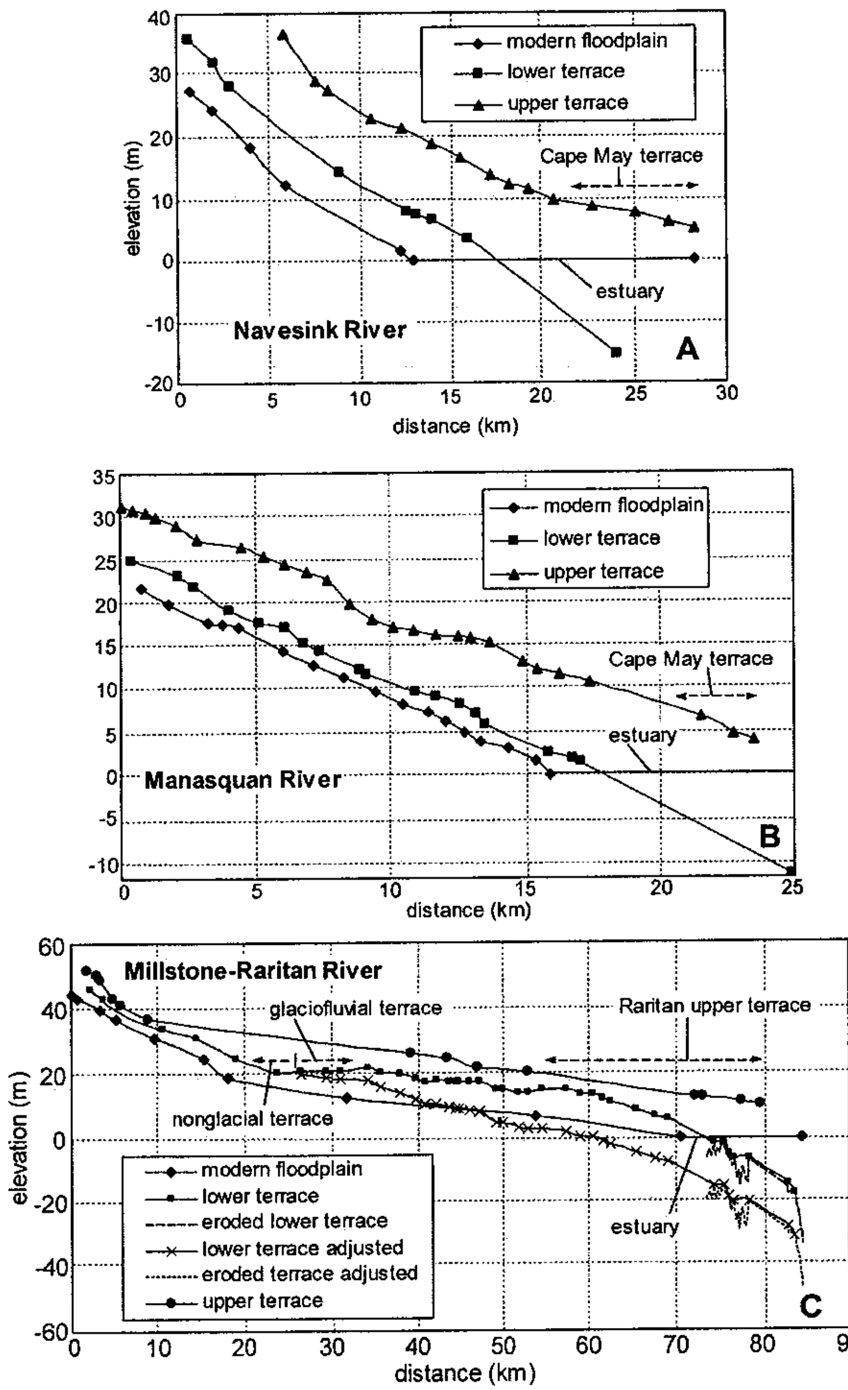

Figure 4. Topographic profiles of upper and lower terrace surfaces, showing relaltionship to glaciofluvial terraces in the Millstone valley (C) and to the Cape May 2 marine terrace at the mouths of the Navesink and Manasquan rivers (A and B).

Radiocarbon dates on organic materials from several locations in the Cape May 2 and 3 are all >40 ka (table 1) and are accompanied by warm-temperate pollen, fossil mollusks, and diatoms, indicating a pre-Wisconsinan age (MacClintock and Richards, 1936; MacClintock, 1943; Newell et al., 1995). Mollusks, diatoms, plant fossils, and pollen from the Fish House clay, a bed within the Cape May 1 in the Delaware valley at Pennsauken, also indicates a warm-temperate estuarine setting for that unit (Woolman, 1897; Owens and Minard, 1979; Bogan et al., 1989). Amino-acid racemization is a natural aging process that allows estimation of the age of shells based on the time-dependent degree of alteration of the amino acid leucine, contained in the shell, from one crystal form to another. This technique has been used to date shells from the Cape May. Shells were obtained from two boreholes and an outcrop on the Cape May peninsula and from a sand pit dug into the Cape May 1 and Cape May 2 near the mouth of the Mullica River (Lacovara, 1997; Wehmiller, 1997; O'Neal and McGeary, 2002). Results indicate a probable 125 ka age for the Cape May 2, which includes samples on the Cape May peninsula from the surface down to an elevation of –50 feet, with 2 or 3 clusters of older shells from beds beneath the Cape May 2, ranging from 200 ka to possibly as old as 2 Ma. These older shells may be from erosional remnants of the Cape May 1, or from non-outcropping deposits of highstands that did not reach modern sea level and were later eroded and buried by the Cape May 2. Of particular interest with regard to the Cape May 1 deposits is the sea-level highstand around 400 ka, during oxygen-isotope stage 11. Global records indicate that sea level at this time was 60-70 feet higher than modern sea level, approximating the maximum height of the Cape May 1 terraces (O'Neal and McGeary, 2002). Parts of the upper terrace deposits may also day from this highstand.

The elevation range of the Cape May 2 terrace agrees favorably with estimates of global sea level during the Sangamonian highstand (125 ka), which are about 20 feet above modern sea level. The lower Cape May 3 terrace was most likely formed shortly after the Cape May 2 as sea level fell from the 125 ka maximum. Newell et al. (1995) consider the Cape May 3 to be a highstand deposit of middle Wisconsinan age (35-80 ka), based in part on a radiocarbon date of 34,480 yrs B. P. on peaty silt that they consider to be within the Cape May 3 beneath late Wisconsinan colluvium (table 1). The associated pollen, however, indicates cool-temperate climate like that during deposition of the lower terrace deposits, and it is more likely that the silt is an alluvial deposit within or beneath the colluvium, not an estuarine or marine deposit. Middle and early Wisconsinan sea levels higher than modern sea level have not been documented on other tectonically stable coasts, and are unlikely here.

Lower Terrace Deposits

During the period of lowered sea level in the early and middle Wisconsinan, after the Sangamonian interglacial, streams incised as much as 50 feet into the upper terraces and Cape May deposits. The lower terrace deposits were laid down in these incised valleys between approximately 35 and 15 ka, based on radiocarbon dates within and at the base of the lower-terrace deposits and at the base of overlying alluvial deposits (table 1). The radiocarbon dates indicate that the lower terraces were deposited in large part during the maximum extent of the late Wisconsinan glacier. This glacier reached its southernmost position on the Atlantic coast at Perth Amboy. It is bracketed by radiocarbon dates to between 21 and 18 ka in New Jersey and Long Island (Stone and Borns, 1986). Thermokarst basins and cryoturbation features, which are ubiquitous on the Cape May and upper terraces and older landsurfaces in the NJCP (Wolfe, 1953; French and Demitroff, 2001), are absent from the lower terraces. This pattern further indicates that the lower terraces were being actively deposited during the last glacial maximum, while cryogenic structures were forming on older surfaces.

The lower terrace deposits are lithically similar to the upper terrace sediments. In some valleys in the inner Coastal Plain, lower-terrace sands are more glauconitic than the upper terrace deposits because incision between the two terraces exposed glauconite-rich formations. During deposition of the upper terraces, the glauconitic formations were covered by quartz sand of the Kirkwood Formation. As with the upper terraces, deposits of colluvium collected in aprons at the foot of some steep slopes adjoining the lower terraces and modern floodplains. These deposits are thin (<10 feet thick) and generally of limited extent.

The lower terraces are much smaller and narrower than the upper terraces and in many valleys have been nearly completely eroded during incision and widening of the modern floodplain. One exception is the South River valley in Middlesex County, where the lower terrace is extensive and the upper terrace is absent. The South River is a tributary to the lower Raritan, which was cut when the pre-late Wisconsinan Raritan valley was blocked by glacial deposits during the late Wisconsinan glaciation, diverting the Raritan southeastward from Bound Brook to Perth Amboy (fig. 3d) (Stanford, 1993). The South River valley was incised, broadened, and extended headward to the south as it adjusted to the newly deepened Raritan valley at its mouth. Lower-terrace deposits were laid down in this new valley. Upper terraces are absent because the valley did not exist in its present form before the late Wisconsinan.

Glaciofluvial Deposits

In the Delaware and Millstone valleys, lower terraces grade to a late Wisconsinan glaciofluvial terrace (fig 4). This glaciofluvial deposit extends down the Delaware valley from the terminal moraine near Belvidere in Warren County. It is the only glacial deposit in the NJCP, outside of the glaciated extreme northeast corner of the NJCP around Perth Amboy. It is composed chiefly of a lithic sand matrix and pebble-to-cobble gravel of gray and red sedimentary rock from the Appalachians and Newark Basin, with some Precambrian gneiss from the Reading Prong and New Jersey Highlands. At and south of Trenton, it includes much quartz and quartzite supplied from tributary streams draining from the Coastal Plain. The gravel clasts are unweathered, except for carbonate-rock clasts, which are generally decomposed. The terrace surface is 50-60 feet above the modern floodplain of the Delaware River at and north of the Trenton area, and gradually descends from an altitude of 55 feet at Trenton to about sea level in the Burlington area. South of there, it occurs only as erosional remnants in the subsurface beneath Holocene deposits in the Delaware estuary. In the Delaware valley at and north of Trenton the glaciofluvial terrace has been extensively eroded in postglacial time by the Delaware River and is preserved intact only in isolated remnants along the valley walls. A postglacial stream terrace, which is largely a sand-capped strath cut into the glaciofluvial gravel, occupies most of the valley bottom, with a surface about 20 feet above the narrow modern floodplain.

The glaciofluvial deposit has long been informally termed the "Trenton gravel" (Cook, 1880). Salisbury and Knapp (in Bascom et al., 1909a), while recognizing its glacial origin, included the Trenton gravel in the Cape May Formation. Owens and Minard (1979) renamed the deposit the "Spring Lake" and "Van Sciver Lake" beds, for locations in the uneroded terrace just south of Trenton (Spring Lake) and in the postglacial stream terrace near Morrisville, Pennsylvania (Van Sciver Lake). At Van Sciver Lake, gravel pits have been dug through the thin postglacial stream-terrace sand into the underlying Trenton gravel. Owens and Minard (1979) considered the Spring Lake and Van Sciver Lake beds to be fluvial-estuarine deposits of Sangamonian age, based in part on a mistaken down-valley correlation to the Fish House clay in the Cape May 1 at Pennsauken, which is a much older deposit. This usage is partially continued by Newell et al. (2000), owing to miscorrelation of the postglacial terrace at Van Sciver Lake to the glaciofluvial terrace upvalley from Trenton. Because of these miscorrelations, the terms "Spring Lake beds" and "Van Sciver beds" are not used by the New Jersey Geological Survey.

At Trenton the glaciofluvial plain broadens and part of it extends northeastward across a low divide to the Millstone valley at Princeton, and continues northward down the Millstone to the Raritan. This unusual geometry reflects the influence of glacioisostatic depression during the late Wisconsinan maximum, which tilted the landsurface down to the northeast sufficiently to partially divert meltwater from the Delaware into the Millstone basin (Stanford, 1993).

Eolian Deposits

Windblown sand and silt are another relict of glacial climate in the NJCP. These deposits are generally of small extent and widely scattered (fig. 5). They include sheets of silt and very fine sand, mostly less than 3 feet thick, and fine-to-medium sand in dunefields up to 15 feet thick. They are located chiefly on valley sides and uplands east and southeast from broad expanses of lower-terrace and glaciofluvial-terrace deposits. This geometry indicates that they were blown from alluviating, unvegetated lower terraces by westerly winds and are of the same late Wisconsinan age as the lower terraces. Other, smaller dune deposits occur on outcrops of sandy Coastal Plain formations, primarily the Kirkwood and Cohansey. These dunes are shaped from the underlying sands, with little or no transport, and may have formed when vegetation was removed, perhaps by fire or storms, and patches of the formations were exposed to wind shaping.

Modern Floodplains and Estuaries

Pollen records indicate that boreal spruce-pine forest had largely replaced the mixed tundra grassland-patchy boreal woodland present in the NJCP at glacial maxima by about 13 ka (Watts, 1979; Russell and Stanford, 2000). This reforestation, and melting of permafrost, reduced surface runoff, slope erosion, and alluviation in valleys. With sea level still well below interglacial level, streams incised into the stabilized landscape. Channels were cut into

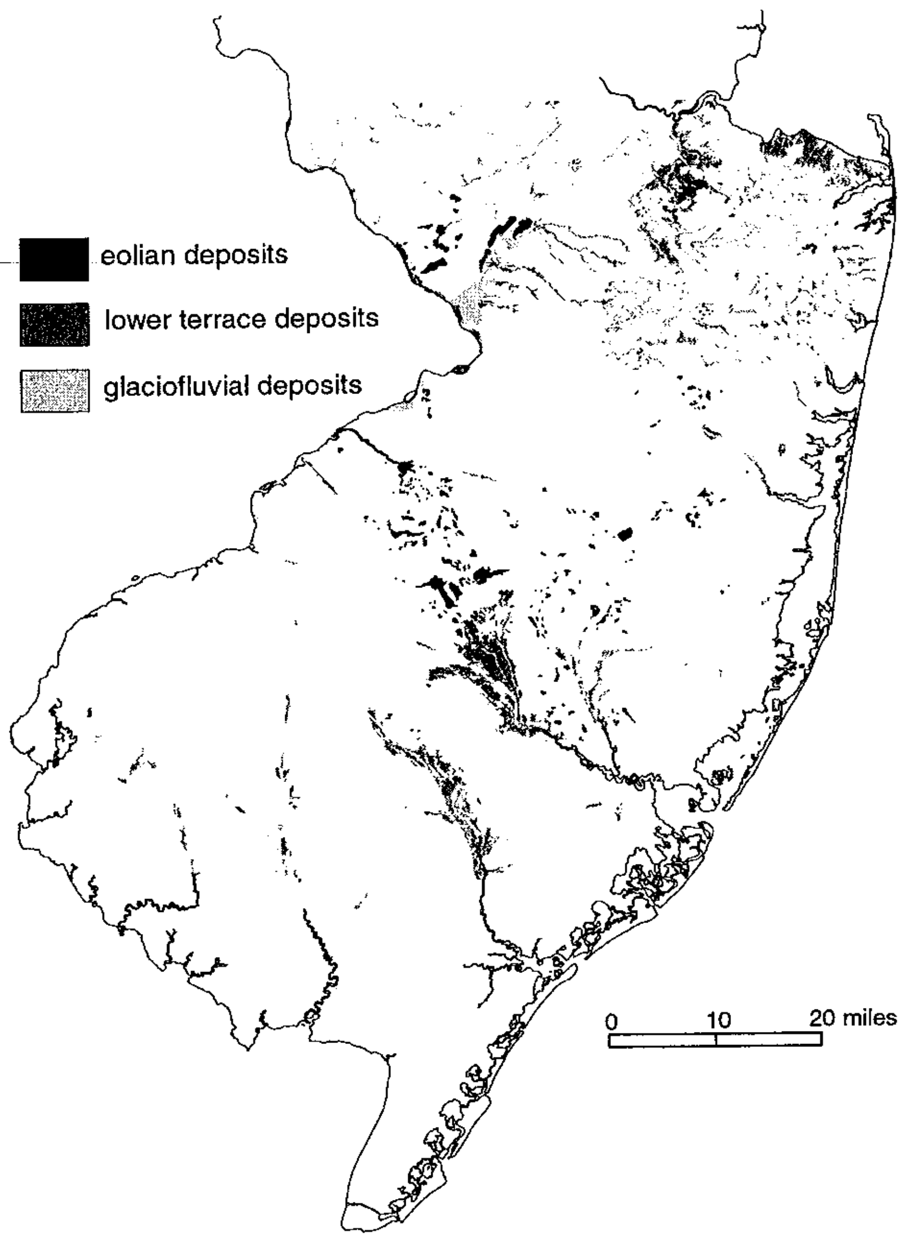

Figure 5. Extent of eolian, lower-terrace, and glaciofluvial deposits. From Newell et al. (2000), with revisions. Glaciofluvial deposits are also present beneath Holocene estuarine sediment in the Delaware valley south of the Burlington area, and in the Raritan estuary and bay.

the lower terraces and the glaciofluvial terrace in the Delaware valley. Incision was generally less than 20 feet in the inland parts of the NJCP but was as much as 100 feet in the lower Delaware and Raritan valleys, in areas now largely buried by estuarine deposits. The Delaware and Raritan continued to carry large volumes of glacial meltwater draining from glacial lakes north of the terminal moraine, after most glaciofluvial deposition has ceased. This added discharge enhanced incision in those valleys.

Rising postglacial sea level entered the deeply incised Raritan and Delaware valleys as early as 12 ka (Owens et al., 1974; Fletcher et al., 1990; Stanford et al., 2002). It has continued to rise since then (Psuty, 1986), with the rate of rise slowing significantly at about 7.5 ka, when the Laurentide ice sheet had completely melted. This rise has resulted in the deposition of estuarine, salt-marsh, bay, beach, and nearshore marine deposits along the coast and lower reaches of river valleys. These deposits are as much as 250 feet thick where they fill and bury former incised late Wisconsinan channels.

Inland, rising sea level raised fluvial baselevels and largely ended valley incision. Alluviation and floodplain development followed. Radiocarbon dates (table 1) indicate that floodplain deposition began before 12-10 ka. Floodplain deposits are generally less than 10 feet thick and include massive to plane-bedded to low-angle cross-bedded sand, silt, and pebble gravel. Peat and organic silt and clay are common on floodplain margins and headwater areas where groundwater seepage from adjacent uplands maintains saturated conditions. Peat is also a dominant component of floodplain deposits in the Pine Barrens, where there is little to no upland surface runoff owing to the high permeability of the Kirkwood and Cohansey sands. Thus, little sediment enters streams. Instead, water reaches streams by seepage of groundwater from the bottom and margins of the floodplains. This constant seepage maintains saturated conditions in the floodplain and allows peat to accumulate. At low elevations in the valleys draining to the Atlantic, sea-level rise has led to a concomitant rise in the water table, which in turn has led to vertical accretion of peat in the floodplains over the past 10 ka, to a thickness of as much as 8 feet (Buell, 1970; Watts, 1979; Stanford, 2000).

Continued rise of sea level, currently proceeding at a rate of about 0.1 inch per year, will cause salt marshes to encroach further inland, covering freshwater wetlands and floodplains. Barrier islands, spits, and mainland beaches will also move landward. If human-caused global warming accelerates the rise of sea level, it is possible that the West Antarctic ice sheet, which is particularly sensitive to sea-level change because it is buttressed by marine ice shelves, could decay rapidly. This decay would raise global sea level by 15-20 feet, returning the sea to the Sangamonian highstand level of 125 ka, when the West Antarctic ice sheet may last have decayed. The extent of the Cape May 2 (fig. 3d), which was deposited at the 125 ka highstand, approximates what the coastline of New Jersey would look like if this were to happen.

SEA LEVEL AND CLIMATE CHANGE

Figure 6 summarizes the depositional and erosional events in the NJCP over the past 10 m.y., and figure 7 correlates these deposits and erosional periods to the global sea-level

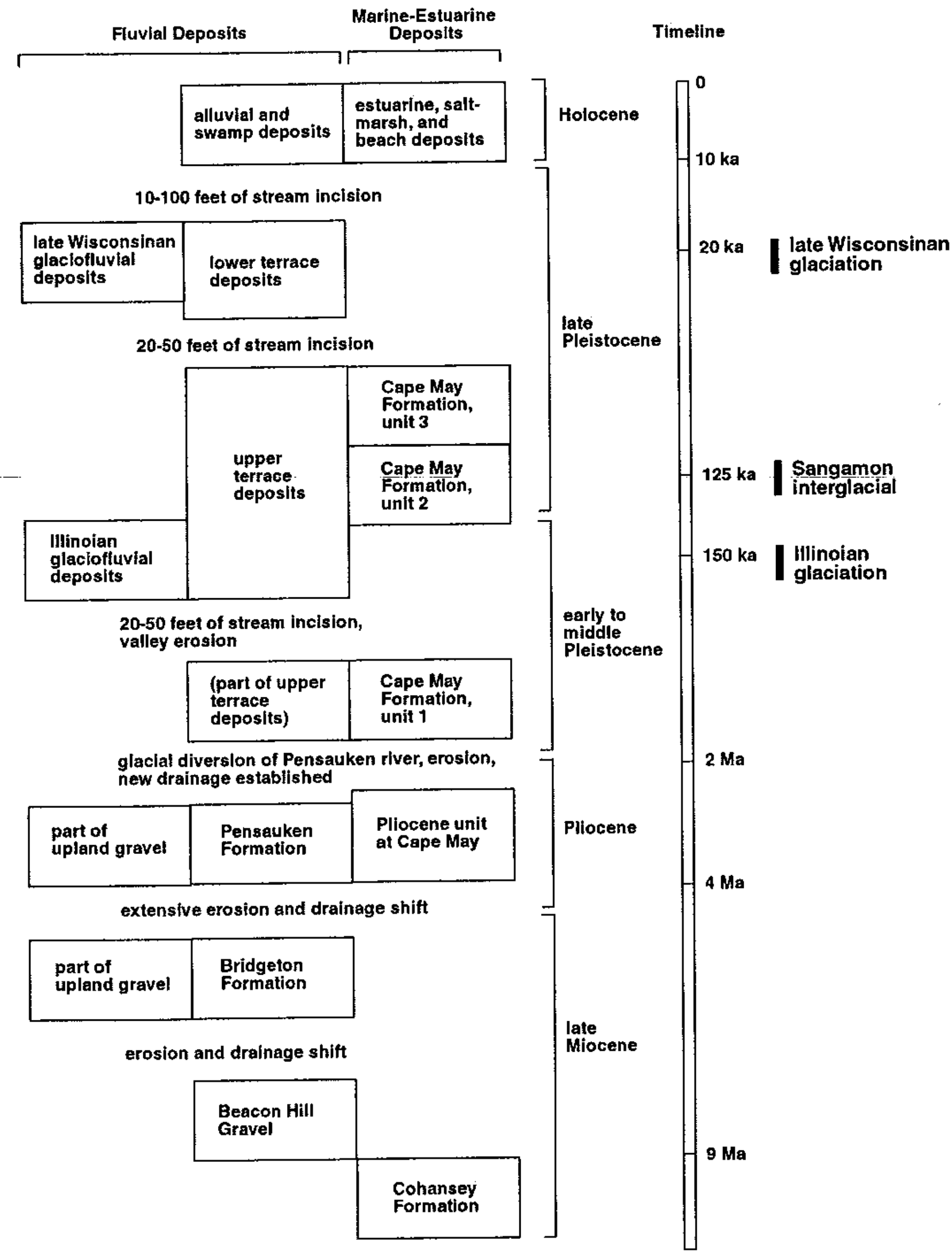

Figure 6. Correlation of deposits and erosional events for the New Jersey Coastal Plain.

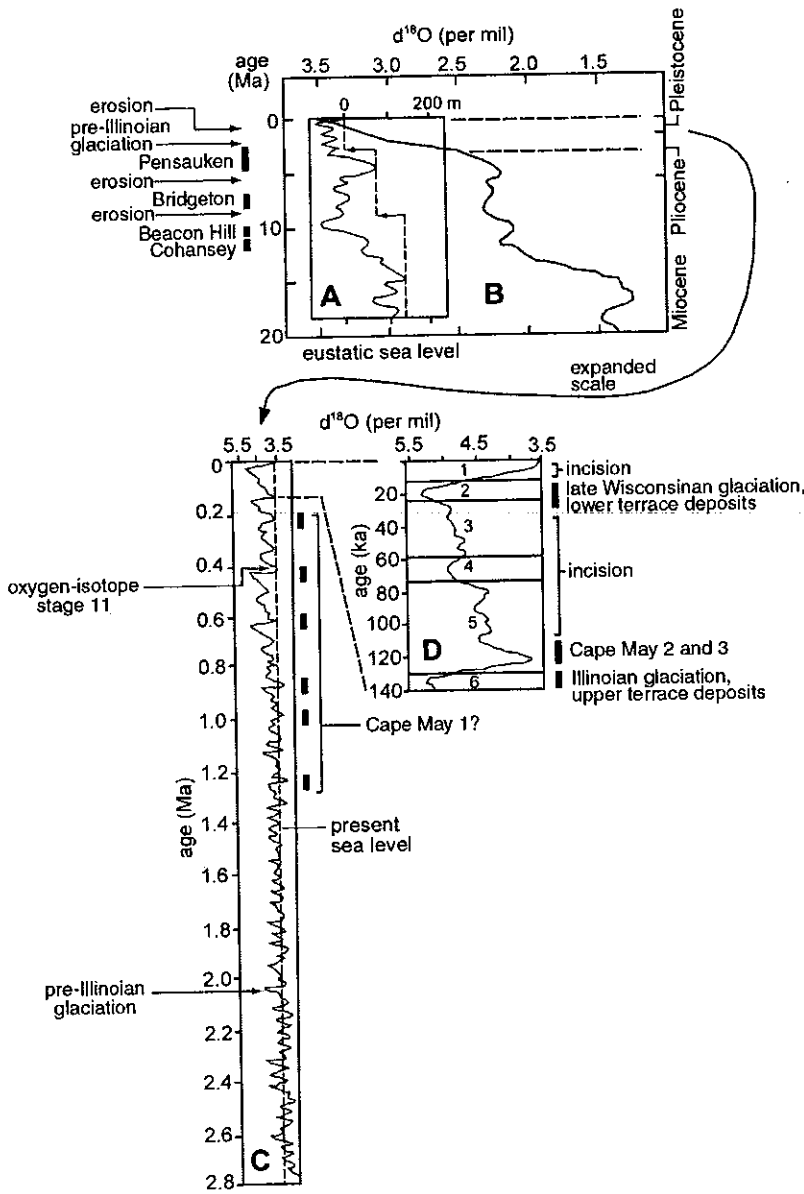

Figure 7. Surficial deposits and erosional periods in the New Jersey Coastal Plain correlated to the eustatic (A) and marine oxygen isotope (B, C, D) records. Arrows and dashed lines onA show permanent eustatic declines related to Antarctic and northern hemisphere ice-sheet growth in the late Miocene and late Pliocene, respectively. These declines correlate to the erosional periods between deposition of the Beacon Hill, Bridgeton, and Pensauken formations., and the erosional period following diversion of the Pensauken river during the pre-Illinoian glaciation. Expanded oxygen-isotope record C shows declining sea-level trend over the past 2.5 m.y. due to progressive growth of northern hemisphere ice sheets, with repeated Laurentide glaciation producing the large-amplitude variations after 800 ka. Oxygen-isotope curve D shows the last interglacial-glacial cycle. Cape May 2 and 3 deposits are laid down during the brief stage 5 interglacial highstand; terrace deposits are laid down chiefly during cold periods (stages 6 and 2) corresponding to glacial advances. Incision occurs before and after cold maxima, when sea-level is still low compared to interglacials. Curve A is from Haq et al. (1987), curve B is from Miller et al. (1987), curves C and D are from Raymo (1992).

record. Curve A on figure 7 shows estimated global sea level for the past 20 m.y. (in meters relative to modern sea level) based on vertical shifts in seismically imaged onlap patterns of continental-margin strata (Haq et al., 1987). Curves B, C, and D show, in increasing resolution, the oxygen-isotope composition of ocean water as recorded in the shells of benthic foraminifera recovered from test holes drilled through ocean-bottom sediments. Curve B (Miller et al., 1987) is smoothed to remove frequencies of < 1 m.y. During glacial expansion, water evaporated from the ocean falls as snow on the accumulation areas of glaciers and is stored as glacial ice. The evaporated water is enriched in ^{16}O, the lighter isotope of oxygen. Thus, seawater becomes enriched in ^{18}O, the heavier isotope, during periods of glacier growth. These periods are marked on figure 7 by leftward swings of the curve. The large enrichments between 15-10 Ma and since 5 Ma on curve B record net ice sheet expansion in Antarctica and in the northern hemisphere, respectively. Ice-sheet expansion causes global sea-level decline, and the oxygen-isotope curve is a proxy for global sea level. The overall similarity of curves A and B demonstrates this link.

The actual magnitude of sea-level change is more difficult to determine. The onlap method is inexact because younger strata may erode upper parts of earlier strata. The oxygen-isotope composition of shells depends on temperature and salinity in addition to the composition of the seawater, although these effects can be corrected by using the ratio of Mg/Ca in the shell. Using this method, the enrichment between 15-10 Ma may represent between 60 and 100 m of sea-level decline (Lear et al., 2000). The glacial-interglacial swings over the past 800 ka (curves C and D; Raymo, 1992) represent 120-150 m of sea-level change

The transition from inner-shelf sedimentation (Cohansey Formation) to fluvial deposition on the exposed shelf (Beacon Hill Gravel) to a lower, inset fluvial plain (Bridgeton Formation) between 11-8? Ma correlates to the tail end of the middle-to-late Miocene sea-level decline. The full depth of incision from the Beacon Hill to the base of the Pensauken, which includes all the late Miocene incision to about 4 or 5 Ma, is about 90 m, within the range of sea-level decline estimated from the oxygen-isotope enrichment.

Aggradation of the Pensauken fluvial plain, and the cutting of straths and lowlands graded to this plain, corresponds to a period of high sea level around 3.5 Ma. This highstand shows up on both the onlap and oxygen-isotope curves, and corresponds to a period of climatic warmth and high sea level recorded in marginal marine deposits around the Atlantic (for example, Dowsett and Cronin, 1990). Elevated sea level at this time may be due to partial melting of the Antarctic ice sheet, although the extent of this melting is uncertain (see, for example, Kennett and Hodell, 1995).

Following this highstand, a trend toward colder climate in the Arctic initiated growth of ice sheets in the northern hemisphere, starting around 2.5 Ma. The first Laurentide ice sheet reached into the midcontinent, and possibly into New Jersey, at around 2 Ma. The oxygen-isotope curve shows that Arctic ice volume was sufficient to keep sea level lower than present sea level for the vast majority of the past 2 m.y. (fig. 7, curve C). This lowered sea level, in combination with pre-Illinoian glacial diversion of the Pensauken river, has favored incision in the NJCP during most of the Pleistocene. Widespread deposition has occurred only at peak interglacials, when estuarine and marginal marine deposits are laid down along the coast and lower reaches of river valleys, and during peak glacials, when permafrost increases surface runoff and valley alluviation. Both these peak periods are brief (<20 k.y.) in New Jersey, based on radiocarbon dates bracketing the lower terrace deposits and late Wisconsinan glaciation, and based on the oxygen-isotope record (fig. 7, curve C), which shows only brief periods of higher-than-present sea level.

In lowland areas across the eastern United States, permafrost features are restricted to a belt less than 100 miles wide south of the glacial border (Washburn, 1973). This pattern suggests that close proximity to the ice is required to generate permafrost at these latitudes. There is evidence for only three glaciations in New Jersey, the pre-Illinoian, possibly at 2 Ma, a probable Illinoian glaciation at 150 ka, and late Wisconsinan at 20 ka. At other times the glacier margin was likely no farther south than the Allegheny Plateau-Catskill Mountains, which formed a significant topographic barrier to ice advancing from the Hudson Bay area. Thus, it is possible that tundra vegetation and permafrost were present in New Jersey only twice for brief periods (<20 k.y.): once at the time of the Illinoian maximum and once at the time of the late Wisconsinan maximum. Much of the landsurface present at the time of pre-Illinoian glaciation has been eroded, erasing any periglacial features that may have formed at that time. Thus, the upper and lower terrace deposits, corresponding to the Illinoian and late Wisconsinan glacial maxima, respectively, and correlative colluvial deposits at the base of steep slopes, graded to the upper and lower terraces, may be anomalous as a record of Pleistocene geomorphic activity in the NJCP. The ubiquity of thermokarst basins and other periglacial features on flat or gently sloping surfaces older than the lower terraces demonstrates that most of the landscape was stable and non-eroding during glacial maxima, with only the steeper slopes shedding sediment. These observations differ from the model of widespread "badland"-style erosion during lengthy cold periods advocated for the NJCP by Newell et al. (1995, 2000), which is difficult to reconcile with the evidence for widespread surface stability.

Under the forested, cool-temperate climate conditions, with lower-than-present sea level, that prevailed during the vast majority of the Pleistocene, the dominant geomorphic processes were likely slow incision by non-alluviating streams and gradual valley widening from erosion of valley sides by groundwater seepage at the base of uplands and margins of floodplains. These groundwater-based processes are the dominant ones today in the NJCP.

CONCLUSIONS

The surficial deposits and landforms of the NJCP record the response of a low-relief passive-margin coastal terrain, of predominantly forested, temperate climate, to the large-amplitude glacially driven sea-level changes of the past 10 m.y. The marginal-marine Cohansey Formation and fluvial Beacon Hill Gravel and Bridgeton Formation record the transition from inner shelf to regressive fluvial plain to incised fluvial plain as sea level experienced a long-term decline of 50-100 m between approximately 15 and 7 Ma, in response to Antarctic ice growth. The Pensauken Formation is an aggraded fluvial plain deposited during a sea-level highstand in the mid-Pliocene at about 3.5 Ma. Diversion of the Pensauken river at 2 Ma and subsequent incision and valley widening through the early and middle Pleistocene between about 2 Ma and 150 ka occurred in response to a second period of long-term sea-level decline, driven by the onset and progressive growth of northern hemisphere ice sheets. Coastal sediments of the Cape May Formation (units 2 and 3) were laid down during the Sangamonian interglacial sea-level highstand at 125 ka and during at least one earlier, higher interglacial highstand. Alluvial and colluvial sediment was deposited in valleys at glacial maxima during the Illinoian (150 ka) and late Wisconsinan (20 ka), when permafrost increased surface runoff. Between these relatively brief periods of deposition at glacial and interglacial maxima, stream incision and valley widening prevailed through the Pleistocene.

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A Geography of Spungs and Some Attendant Hydrological Phenomena on the New Jersey Outer Coastal Plain

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ABSTRACT

"Spungs," "cripples," "blue holes," and "savannahs" are archaic and colloquial terms used in the Pine Barrens of southern New Jersey. In modern usage, they ambiguously describe congeneric, aquifer-related, geologic features. Little information is available concerning their geomorphology, their historical significance, or why they are fading from the landscape.

In this paper, four distinct hydrologic features are defined, using terminolgy coined and traditionally employed in the Pinelands. An accounting is given of the origin and development of each term. Based on recent cold-climate geomorphological investigations, a geologic provenance is proposed for each of the described features. Through archeological and historical accounts, the period of their utility is chronicled.

Spungs are enclosed wetland basins, created by deflation under periglacial conditions. They served as oasis-like watering places for wildlife and ambulant peoples over a period of 12,000 years. Cripples are short, broad, and damp, but wooded paleovalleys which lack a modern stream channel and were created by surface wash flowing over permafrost. Blue holes are deep, strong springs, of some antiquity, and are found in river channels, or occasionally on the broad paleovalleys which border a river. Savannahs are grassy, sparsely wooded wetlands which once occupied numerous large paleovalleys in the Pinelands. Since all four features are dependent upon the near-surface watertable for recharge, escalating demand placed upon the region's aquifers imperil what remains of these important wetlands and springs. Anecdotal and archival records affirm that a drop has occurred in the surficial watertable over the last two and a half centuries, with the depletion accelerating since 1930.

INTRODUCTION

The pine region of southern New Jersey is a landscape of subtlety and nuance. There are no grand landmarks, such as lofty mountains or mighty rivers, within its borders. To the casual observer, the region is an uninspiring flatland, a monotonous plain of stunted pine and oak forest. The earliest settlers considered the region a barren desert. Farmers held little interest in the Pinelands' impoverished gravelly and sandy soils.

It was the timber stock, uniquely adapted to the region's inhospitable conditions (i.e., drought, frequent fires, and nutrient-poor soils), that attracted the economic stimulus to inhabit the pines. Demand for lumber, naval stores and charcoal was great during the Colonial period (Boyd, 1991), both for domestic use and for export to England. Forest resource extraction necessitated large teams of woodsman to fell, kiln, prepare and haul the forest products to market.

To the woodsman, slight changes in the landscape were important to note. In a search for forest products, certain environments would prove better than others for exploitation. Vernacular, non-standardized names were coined for specific features (Moonsammy, et al., 1987, p. 13-41), perhaps where water was plentiful, or cedar could be cut, or game was plentiful. To the outsider, the distinctions between certain features would be unclear.

Such is the case for localized hydrologic terms, such as "spungs," "cripples," "blue holes" and "savannahs." As quoted by G.A. Chamberlain (Bisbee, 1971, p.290), "People have grown old and died arguing as to the difference between a cripple and a spung." In a modern context, they are out of place. Their names are archaic, with their original roots long forgotten. The woodcutters, who originated and used these terms with a "sense of place," are now long gone. Lastly, significant changes to the region's groundwater, as well as other anthropogenic factors, have degraded many of the spungs, cripples, blue holes and savannahs beyond recognition.

THE PINE BARRENS

The Pinelands of southern New Jersey consists of gently rolling sandy terrain, extensively covered with pine and oak forest. Numerous rivers, streams and wetlands dissect the sedimentary deposits which underlie this tract. The wetlands support mixed hardwood forests, with some stands of Atlantic White Cedar. Important rare and endangered plants and animals are found within its borders, protected within the 1.1 million-acre Pinelands National Reserve, created and partially funded by Congress for environmental management and land acquisition in 1978 (Pinelands Commission, 1985, 1997). The Pinelands are administered by the Pinelands Commission, which is under the State of New Jersey's jurisdiction.

Geology

The Cohansey and Bridgeton formations underlie much of the land surface. Deposited during the Mid- to Late-Miocene (~15 to 5 million years ago), they consist of marine, marginal-marine, estaurine, deltaic and fluvial deposits. Sands and gravels are common, but smaller deposits of silt, clay, cobbles and boulders are also present (Newell, et al., 2000). Since their deposition, the beds have undergone weathering, and exhibit the effects of prolonged leaching (Owens, et al., 1983, p.35). Tedrow (1986, p.22) states that these surfaces have been modified by slopewash and eolian activity.

Edaphics - How the Sandy Soils Have Influenced Living Things

The earliest explorers first noted the Pine Barrens of southern New Jersey as an arid, inhospitable wasteland. Maps of the 17th and 18th century delineate the region as sandy barren deserts (Holland, 1776; Evans, 1749). In 1722, John Fothergill, the English physician and close friend of Ben Franklin, tells of his "journey through the deserts" of South Jersey's Pinelands, and of the impoverished "few dark people" who inhabited it (Fothergill, 1754). Seven generations later, in 1889, Dr. Isaac Hull Platt (1889, p.5) echoed Fothergill's sentiments. In his address to the American Climatological Association, Platt noted that the dry, sandy, sterile soils of the Pine Belt renders to its inhabitants the most marginal of existence. "His home is a hovel, and his food indigestible, and often insufficient and innutritious"

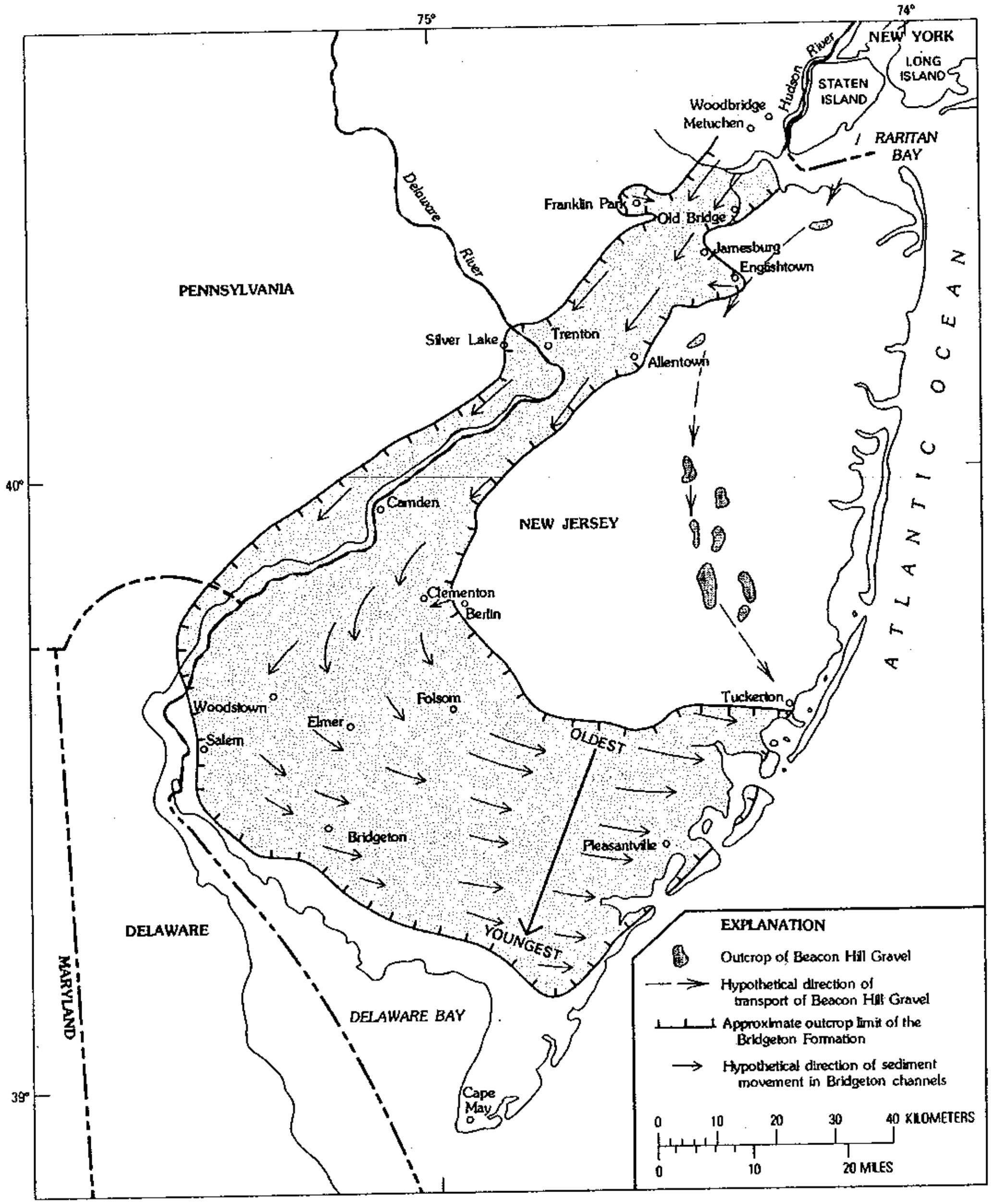


Figure 1
Location map adapted from Owens and Minard (1979, Figures 5 & 13), showing the theoretical flow of the proto-Hudson River during the Miocene. It's believed to have placed the now iron-stained clays, gravels, cobbles, boulders, and angular sand sediments atop the whiter, well-rounded marine-like sediments of the Cohansey Formation.
(A.) First it deposited the Beacon Hill Gravels,
(B.) then it laid down the Bridgeton Formation over the older Cohansey sands. The Cohansey is considered an ancient subdelta plain, which was subjected to wave and tidal action. Modern equivalents are the Nile, Mississippi and Mekong deltas (Rhodehamel, 1979a, p.48).

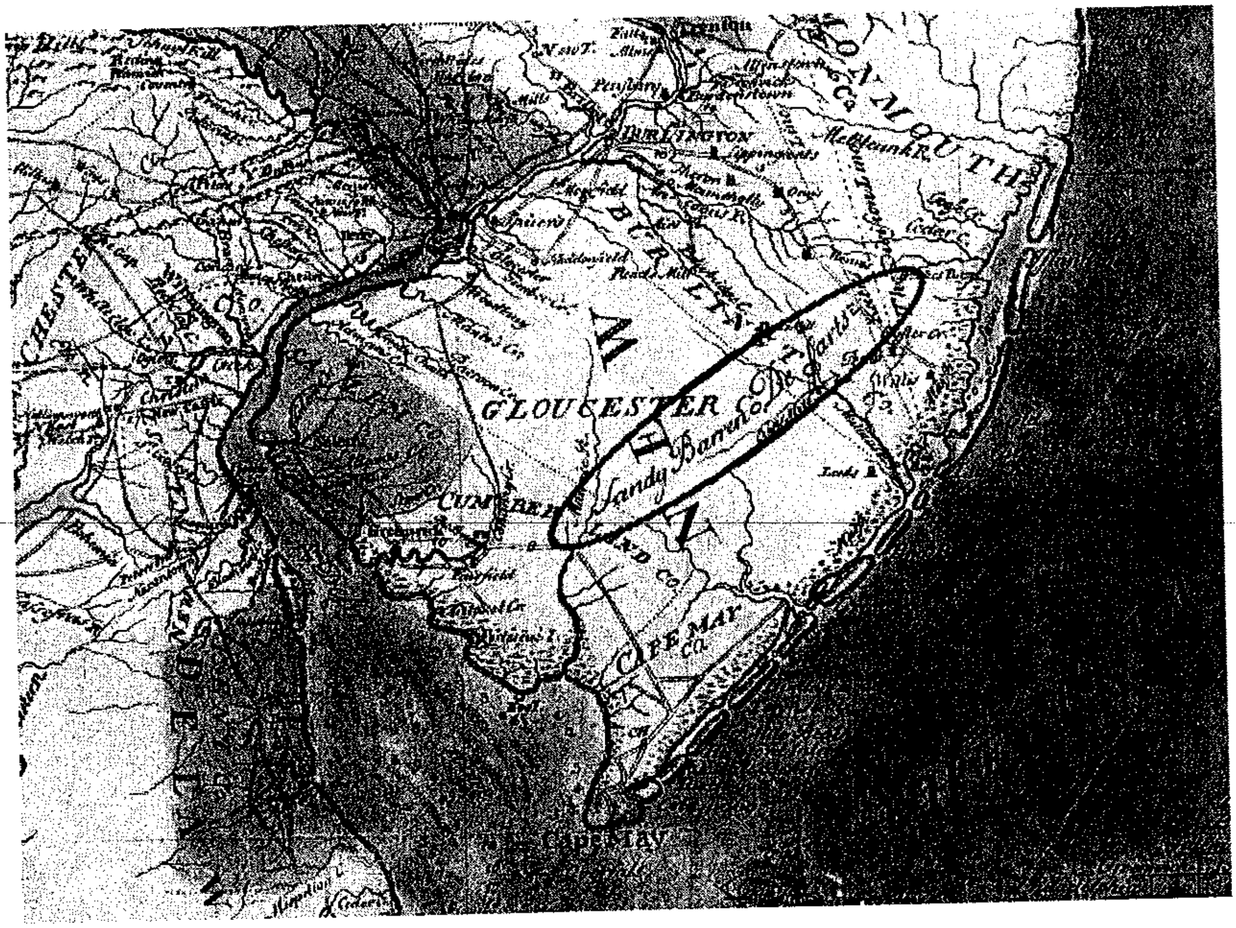


Figure 2
Early mapmakers often designated the outer coastal plain of southern New Jersey as a sandy barren desert (Evans, 1749). Loose cover sands veneer many of the region's surfaces (French and Demitroff, this volume).

Although precipitation is plentiful, averaging more than 45" a year at Hammonton and Indian Mills (Ludlum, 1983), the loose, sandy and gravelly soils allow rainwater to quickly infiltrate, and recharge the shallow aquifer (Rhodehamel, 1979b). Nutrients are leached away, leaving the soil impoverished and considerably acidic. It is the nature of the soil rather than the climate (i.e., the edaphics) which make the Pinelands so drought-prone. The flora and fauna are well adapted to the xeric conditions, and to the frequent forest fires which result (Boyd, 1991, pp. 15-17; Little, 1946).

Shallow Hydrology

In this edaphic desert, the groundwater is never far beneath the surface, averaging 3-6 feet in depth. The Cohansey aquifer has excellent hydraulic characteristics, and holds prodigious quantities of pristine water. Wherever the porous sands are punctured, wetlands and streams appear (Rhodehamel, 1979b). Most stream flow is the result of groundwater seepage from the Cohansey aquifer. Runoff is negligible in the Pinelands (Vermeule, 1894; Rhodehamel, 1979b). As with other Pine Barrens streams and wetlands, the numerous depressions and pools that are known as "spungs" also derive their recharge by intersecting the water table (French and Demitroff, 2001, Fig. 5). This makes them distinct from classical vernal pools (i.e., pools which regularly go completely dry). The Pinelands ponds are entirely groundwater-replenished, whereas a typical vernal pool fills primarily from runoff, although often supplemented by groundwater (Kenney and Burne, 2001). Spungs are open systems, dependent upon the groundwater seeping through the enclosing sandy and gravelly sediments. They are not perched systems that are filled by interception.

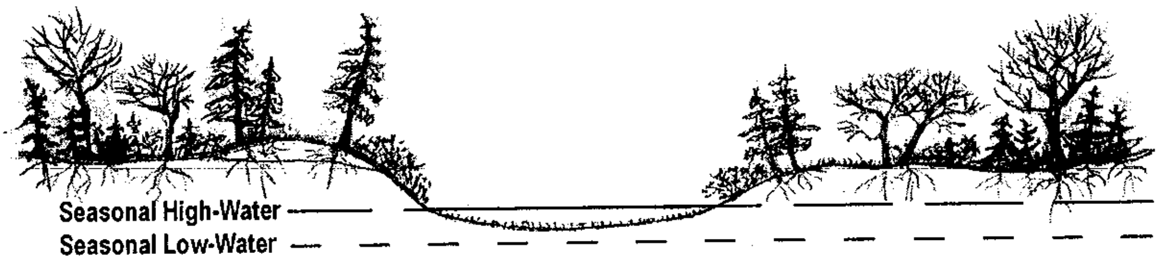


Figure 3
Diagram of a spung showing the seasonal relationship between the ground water table and the apparent water level exhibited within it. When the surrounding groundwater drops below the pond bottom, the spung is waterless. Dashed lines show seasonal high and seasonal low water tables. Drawing by P. Demitroff.

SPUNGS OF THE PINE BARRENS

Etymology of Spung

The use of the word spung appears to be confined to the Pine Barrens of New Jersey (Bates and Jackson, 1987, p 636). The pronunciation is widely accepted to rhyme with the word "rung". However, the way it is written is a bit more problematic. Maps and deeds often used the spelling "spung" during the 17th and 18th century, but by the 19th and 20th century, "spong" is often preferred.

The *Oxford English Dictionary* (1971) describes both spellings. A "spung" is defined as a purse or fob, a small "pocket" formerly made in the waistband of breeches, for carrying a watch, money, or other valuables. It featured a composition of wash-leather or stout lining material (Cunningham, 1971). The verb "to spung" someone is to pickpocket them (*Oxford English Dictionary*, CDROM, second edition, version 3.0). In contrast, the *Oxford English Dictionary* (1971) defines a "spong" as a long narrow piece or strip of land, which does not adequately describe the features as observed in South Jersey. Environmental consultant Jack McCormick (1979, p.231) delineates spungs as "nearly circular depressions that are scattered through the region". The author agrees with both McCormick's spelling and description of a spung, which McCormick likely learned from his mentor, Silas Little, the eminent Pinelands forester (Gordon, Ted, pers. comm., 2002).

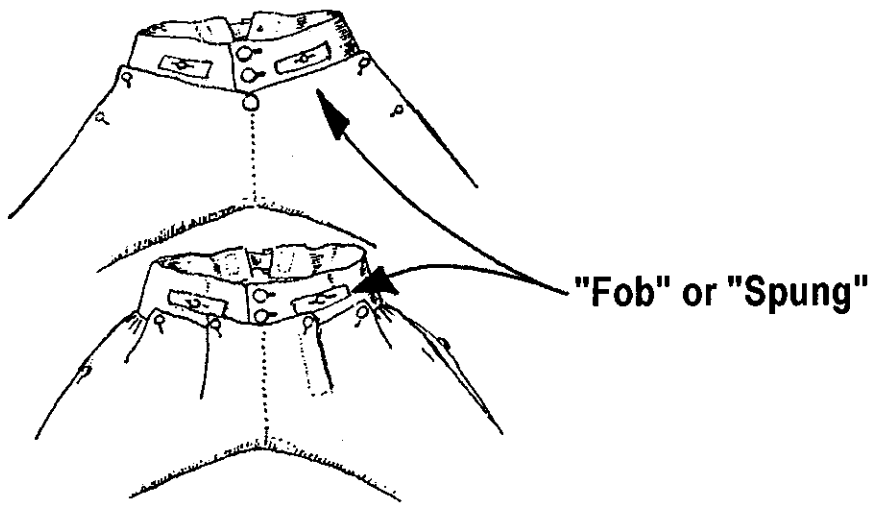


Figure 4
An illustration of breeches, c.1700-1750, (adapted from Cunningham, 1971, Figure 18) showing the fob pockets in the waistband (adapted from Cunningham,1971). A "spung" is a fob pocket, a part of the breeches construction. It is a small (~ 3 inches across) horizontal pocket, set in just under the waist (Lister, Assistant Curator, Museum of London, pers. comm., 2003). Breeches were popular from the 16th through the early 19th century, and were worn by men from all socioeconomic levels (Colonial Williamsburg Foundation, 2003). Illustrations of the spung pocket are difficult to find for two reasons: the coat and the waistcoat cover this portion of the male figure. Displaying this portion of the garment would have involved a very informal pose contrary to the social mores of the time (Gossman, Design and Technology, Furman University, pers. comm., 2003).

Pinelands spungs are simply small enclosed "pockets", often charged with water for at least part of the year. In Ocean County, the term "pocket bog" is often used to describe a spung. A grizzled old denizen of the Pines once rationalized that this appellation refers to the ease of preparation that pocket bogs afforded for cranberry culture. As spungs are often bare of woody vegetation, one could sprig out a planting with minimal investment and without the backbreaking toil of turfing. Consequently, the cultivator could pocket the savings in labor (Emery, pers. comm, 2002). However, the author believes "pocket bog" is a direct translation of the now obscure word "spung". Other monikers include "prim ponds," "goose ponds" (Beers, 1872a), "elephant" or "hog wallows," "holes," "winter ponds," "watering places," "heath ponds" (White, 1885), "dry ponds" (McCormick, 1970) and "fens" (Collins and Anderson, 1994).

Epithets of Spungs

Many of these ponds had descriptive names, most of which are long forgotten. Ancient deeds, surveys and maps often employed these features as important markers to designate property boundaries. Old road returns are excellent resources, since early byways were often modified Indian trails, on their way from pond to pond. Gunning clubs are also good sources for uncovering a spung's name, as they remain an important resource for hunters to this day. Local residents, usually much advanced in years, often remember the names of their childhood fishing, skating and swimming holes, now consigned to oblivion.

Hunting themes are common to the pools. The Bears Hole (Pancoast Mill), Crane Pond (Mays Landing), Big Goose Pond (Egg Harbor City) and the Hog Holes (Millville) are all obviously named for the game which could be stalked there. Others are more archaic in name signification, like Sound Pond (Hammonton), an obsolete hunting term for a spring or pool of water (*Oxford English Dictionary*, 1971). Spungs can also have geographically descriptive names like Desolation Pond (Nesco), Round Pond (Milmay) and the Punch Bowl (Milmay). Blue Bent Pond in Mizpah was a perfectly circular pond. The "bent" is derived from the curved portion of a round form, now rare in usage (*Oxford English Dictionary*, 1971).

Another theme is to identify their importance as way-stops for thirsty travelers. Horse Break Pond (Buena), Jacob's Well (Williamstown), Watering Place Pond (Warren Grove), and the Oasis (New Italy) are good examples of resting places. Others are not so obvious, like Wall Pond at the head of Tarkiln Branch (near Belleplain), which may relate to an obsolete nautical term. The pine woods surrounding this pond were kilned for naval stores production, essential materials for the local shipyards and exportation to England. To "lie at the wall" is to rest at dock (*Oxford English Dictionary*, 1971), and Wall Pond did serve as a resting spot.

Few places in the Pines could provide as good access to potable water as a spung. The small, misfit streams in this region are located in broad swamps which are adjacent to the waterways. These are interpreted as paleovalleys (Farrell, et.al. 1985; Newell and Wyckoff, 1988; French and Demitroff, in this volume) that relate back to Pleistocene conditions. Getting your livestock to a stream also had its perils. Animals had to negotiate the wide swamp to reach open water. Main Road in Vineland was previously known as Horse Bridge Road, in recognition of several horses that permanently mired while trying to cross one such broad floodplain on their way across the Blackwater Branch, sometime prior to 1781 (Ackley, 1929; Hampton, 1917).

Spung Morphography

Many spungs are rounded in outline, while others are irregular. The average depression covers less than an acre, with the largest reaching a hundred acres or more (e.g., Broad Pond, near Elmer; see Mihlasky and Del Sontro, this volume, for examples of spung area measurements). Spungs periodically flank the terraces of many Pinelands streams and rivers, and their nearby uplands. A few are riverine, and those tend to occupy the uppermost channels of smaller streams.

On the higher positions (the interfluves) between rivers, small and medium-sized spungs are inclined to have parabolic bottoms, becoming flatter as depression size increases. Spungs which occupy low positions in the landscape, such as those in broad paleochannels, have remarkably planar floors. This basin morphology could be explained by a permafrost-eolian model. When permafrost conditions existed, higher terrain on the interfluves probably had a low ice content which poorly bonded the enclosing sediments, creating a classic parabolic blowout shape as can be seen in active coastal dunes today (Jungerius and van der Meulen 1989, p. 370, fig. 1).

Low-lying sediments would have a considerably higher moisture content in comparison to the desiccated uplands. Even in arid polar regions, wetlands commonly occupy many low-lying surfaces (Tedrow et al., 1968; Woo and Young, 2003). At our riverine spungs, a near-surface watertable would prevent wind scour from exceeding the coherent surface of such a table. This mechanism is similar to a Stokes surface (Stokes, 1968, Fryberger, Schenk and Krystinik, 1988). The difference is that our low-lying sediments were likely frozen when frigid northwest winds (French and Demitroff, 2001, p.346) eroded the spungs. These flat bottoms are hard packed, or indurated, much like a fragipan, starting at the floor surface.

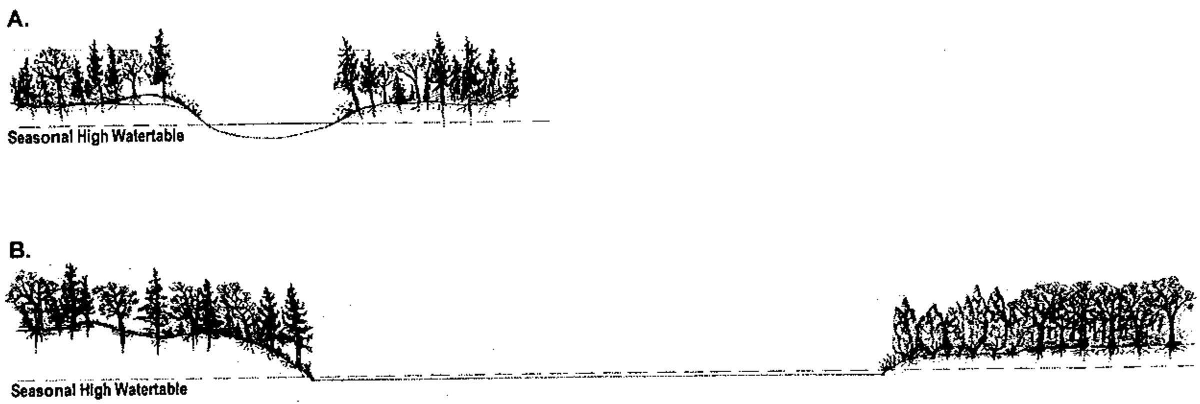


Figure 5
There are two distinct shapes to spung bottoms in the Pine Barrens of southern New Jersey, which could relate to an eolian origin. The first type (A) has a parabolic floor, and is the most common sort. Smaller ponds, which occupy higher and drier positions, almost always exhibit this attribute. Ponds with floors which are flat (B) are usually the larger spungs, and are positioned in lower areas where the groundwater is close to the surface. Drawing by P. Demitroff.

Early Spung Use and Morphology

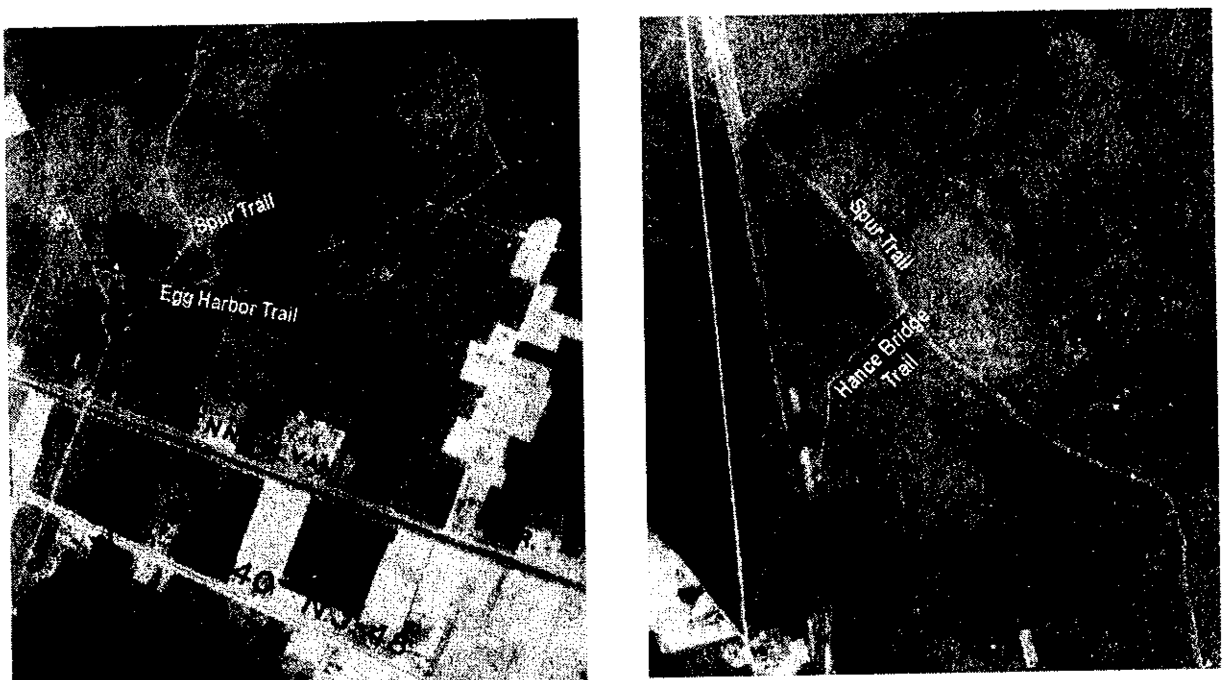
The association of spungs and their use by prehistoric peoples is well established (Bonfiglio and Cresson, 1982; Mounier, 2003), especially as it relates to the depressions found outside the Pine Barrens. Construction activity is lively on the heavier sediments of the Inner Coastal Plain, outside the Pinelands' border. Disturbance due to construction at such sites often necessitates environmental survey work, providing archeologists numerous opportunities to investigate the many depressions described by Bonfiglio, Cresson, and Mounier. However, the spungs of the Pine Barrens have been passed over, since the absence of development pressure equates to a lack of economic impetus for their scrutiny (Cresson, pers. comm., 2002).

At variance with the observations of Bonfiglio and Cresson (1982, p. 35, Fig. 6B), the Pinelands spungs tend to be slightly "rimmed" or raised towards the southeast, and do not share the southern and western bias of similar Inner Coastal Plain depressions. It is the author's impression that the utilization of Pinelands watering holes favored the south to southeast positions., related to slightly higher topography found. Farmer Ron Bertonazzi of

New Italy has collected an assortment of precontact artifacts relating to the Archaic (ca. 3,000-8,000 B.P.) and Paleoindian (8,000-12,000 B.P.) periods (Cresson, 2000) from a spung complex known as the Vanaman Thick N' Hole Tract.

Trail Linkage

In the Pinelands, there is a marked association between Indian trails and spungs. Paths linked these watering holes in a chain of way-stops for the earliest inhabitants. For example, numerous such connections are apparent in Buena Vista Township, Atlantic County. The Hance Bridge Trail, within a distance of 4 miles, linked the Oasis spung in the Vanaman Thick N' Hole Tract (New Italy), with the Lummis Swamp spungs (Richland), to an unnamed cluster of spungs above Richland, at St Augustine's Preparatory School. From here, a hunter could continue to the Bears Hole (Pancoast Mill). Alternatively, he could take the Cohansey Trail, which intersects at the unnamed spungs, westward a mile to Horse Break Pond (Buena) and an additional 6 miles to Egg Harbor Pond (Vineland).



A B

Figure 6.
Ancient pathways often intersect on the southeast portion of spungs, where the terrain tends to be slightly raised, and a stalker would be downwind from his prey. Hunters still prefer to set their deer stands on the south and southeast sides for this reason. The Egg Harbor trail traverses the southeastern corner of Horse Break Pond (A), and then divides into several trails as it heads eastward. This is a trail pattern common to many spungs. A mile up, the spur trail intersects the Hance Bridge Trail on the southeastern shore of the Bears Hole (B). Adapted by M. J. Mihalasky and T.S. Del Sontro from Airplane Atlas Sheets, 1930/32, No. 223.

OF CRIPPLES AND SPUNGS

The meaning of "cripple" is confused and frequently associated with "spung" in modern usage, and in the literature (Weygandt 1940, pp.48-50, McPhee 1967, p.61, Bisbee 1971, p.287, p. 290, Bates and Jackson 1987, p.155, 636). It is the opinion of the author that both spungs and cripples are distinct features, each related to cold-climate geomorphology. Unlike the enclosed spungs, cripples have open drainageways. At certain times cripples can behave like rivers (Bates and Jackson, 1987), but lack an incised channel.

The word cripple is derived from the Dutch "kreupelbos" (thicket, brake or underwood) and "kreupelhout" (underwood, undergrowth; Van Wely 1973). According to J.J. de Vris (pers. comm., 2003), Dutch hydrologist from Vrije Universiteit, Netherlands, "kreupelhout" translates to "cripple wood," which are the bushes and underdeveloped low trees that one finds under wet and marshy conditions in Holland today. They usually consist of alder, willow and birch.

Dutch and Swedish colonists settled Southern New Jersey and adjacent Delaware early in the 17th century. They used the term "Creupel" to describe short wetland corridors, a term which later became Anglicized to "Cripple." A very early description (1676) of a cripple relates to a right-of-way for livestock passage in Colonial Delaware. A landowner employed Martin Garretson to make a passageway over a "Valley & Creuple; but could not make ye sd way Sufficient for Cattle to goe over; by Reason of the Rottenness of ye ground, being a Quaking mire wch hath noe foundation for a way..." (Fernow, 1877). This establishes that a cripple occupies a valley, and is wet. A map survey at the Millville Historical Society shows the "lands, swamps and cripples" along the Maurice River (Miller, 1749), again intimating that cripples are stream-like. "Cripple" is frequently employed in land descriptions during the 18th and early 19th centuries in association with swampland.

Thin and discontinuous permafrost occurred at least twice in South Jersey during the Pleistocene (French, Demitroff and Forman 2003, French and Demitroff, this volume). Surface wash from upland catchment areas would create broad, flat drainageways over the impermeable frozen ground. Intermittent streams would have lacked the mechanical energy needed to erode frozen sediments (Kuchukov and Molinovsky, 1983). The active layer would have been fairly shallow in wetland areas, needing more latent heat for thawing than in higher and dryer spots (Woo and Young, 2003, p.297).

When the permafrost thawed, the surrounding "thirsty" sands and gravels of the Pine Barrens intercepted much of this surface flow. What reaches a cripple's catchment today without a permafrost-enhanced perched watertable is insufficient to incise into the paleochannel. Thus, the stream network becomes almost inactive (Vandenberghe and Woo, 2002, p. 491). Cripples bear a semblance to periglacial dells described in the European Pleistocene literature (e.g., French 1996, p.270, 272; Jahn, 1975. p.172, 173). Periglacial dells are short and shallow valleys, often smoothly U-shaped, created under conditions related to prior permafrost action (Hamelin and Cook, 1967). The popular nursery song, "The Farmer in the Dell" takes place in such a valley.

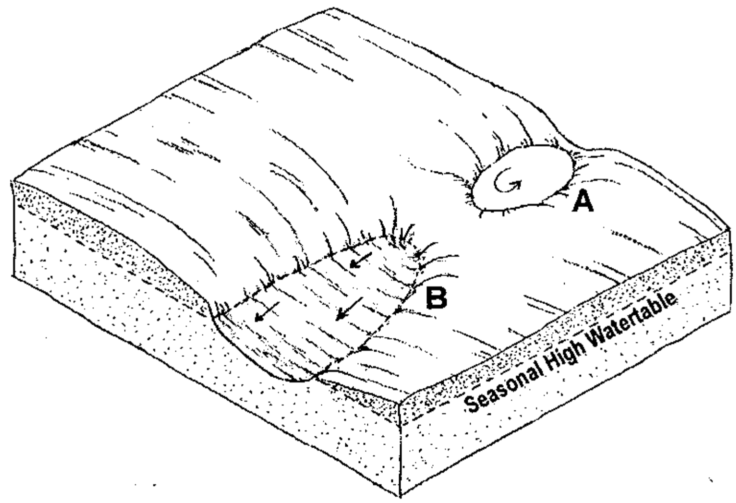


Figure 7
This is a diagram to illustrate some differences between a spung and a cripple. (A) Most spungs are closed systems of the "pocket" type, with internal drainage. Water can overflow the confines under high infiltration phases. (B) Cripples are open systems, involving wide overland drainage (sheet flow) over an abandoned paleochannel under high infiltration phases. Within a short distance, cripples empty into larger drainage systems. Drawing by P. Demitroff.

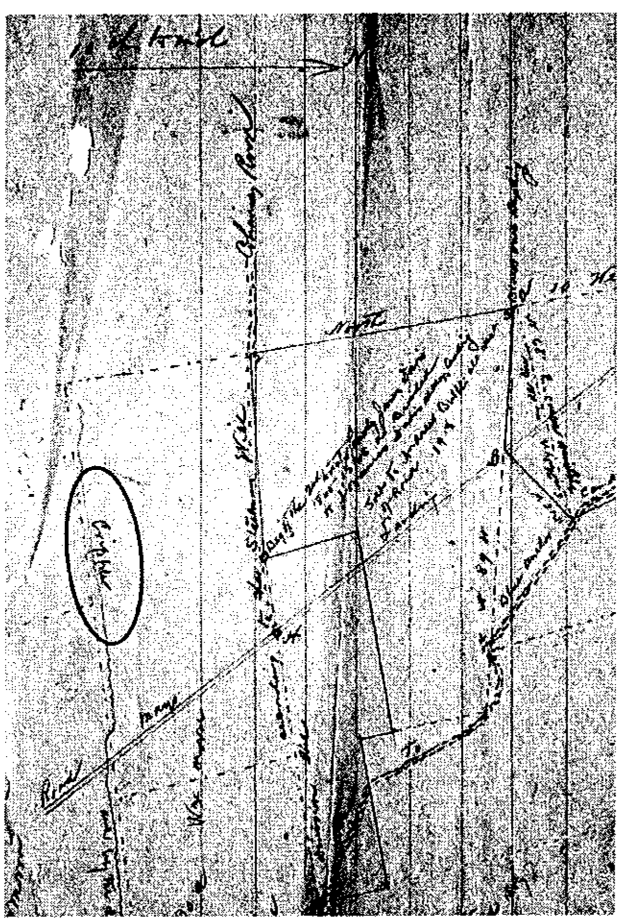


Figure 8.
In this survey, and in others from the 18th and 19th century, a cripple is clearly shown to be elongated and river-like, and are not enclosed and spung-like (Davis Collection #229, c.1867, courtesy of Hamilton Township Historical Society).

OF BLUE HOLES

A number of riverine springs are known as "blue holes" by local inhabitants of the Pine Barrens. Unfortunately, most of these water features are poorly documented in the literature. The author was raised on a farm in the Pine Barrens of Atlantic County, and has heard many tales associated with these springs. Blue holes are often deep, circular cavities occupying the bottom of a streambed. Surcharged by a strong hydrdrostatic head, their sparkling "blue" waters contrast with the surrounding tea-colored "cedar waters" of the Pinelands. Each is as sinister as the next. All are reputed to be bottomless, and all possess dangerous "whirlpools". Young bathers were duly warned not to swim over them, as the icy jets of water emanating from their fathomless depths would cause immediate cramping. You could sink like a stone in the brisk current, or worse yet, be pulled under by the suckhole. Or was it really the fault of the Jersey Devil, abroad from his subterranean lair, hunting for yet another victim?

Only the Inskeeps Blue Hole has received popular attention, beginning with Beck's book, *More Forgotten Towns of Southern New Jersey* (1937, 148-153). Many articles have since told and retold the story which Beck introduced (Ribeiro, 1996, p. 6-8.; Hilton, 1999; McDonald, Walsh, 2003). The site has not received much scientific scrutiny regarding the origin of this phenomenon. In a small paper from December 1958, Genevieve Powell attempted to look at the Inskeeps Blue Hole as a geological problem. She sampled water and soil from the site, and concluded it to be a manmade feature (Powell, 1958).

"Dog Heaven" is another of Atlantic County's blue holes, located on Stony Brook, believed by the author to be the same as the Ingersoll Branch of Absecon Creek. Dog Heaven was just west of where the borders of Pleasantville, Absecon and Egg Harbor Township converge. It, too, was "bottomless", with "whirlpools" and a diabolical inhabitant in its depths. Dogs unfortunate enough to pass its shores would simply fall over dead (Egg Harbor Township Tercentenary Publications Committee, 1964, p.68). The Blue Hole of Newtonville, with its fantastic story of racism, and a mysterious drowning by a suckhole (Bennett, 1999), is a manmade excavation for the nearby bridge embankment. The name is borrowed from a natural blue hole on Three Pond Branch, less than a mile away, just above the Jackson Road bridge. Its lore is replete with a bottomless pit, whirlpools, and the like. "Put anything in it, and you would never see it again" (Huff, Jones, pers. comm., 2000).

Usually, blue holes occur in modern stream channels. They are rarely found on the bordering broad paleochannels. The Inskeeps Blue Hole is an exception. From its source in Berlin, and on to the Blue Hole, the Great Egg Harbor River occupies a broad valley bottom, a channel typical of braided rivers in the high Arctic today (Hamelin and Cook, 1967, p.115; French 1996, Fig 11.8). Polar streams in small basins are usually ephemeral in nature, with nival floods in late spring and little flow by late summer (Woo, 1993; French 1996, Fig. 11.5). In cold semi-arid regions, 30% to 90% of the runoff occurs within a few days or weeks (Vandenberghe and Woo, 2002). Where poorly vegetated, as was probably the case in the Pinelands during the cold period of the Late-Pleistocene, the streams would deposit their large suspended load during this short flood period, further encouraging the braided pattern.

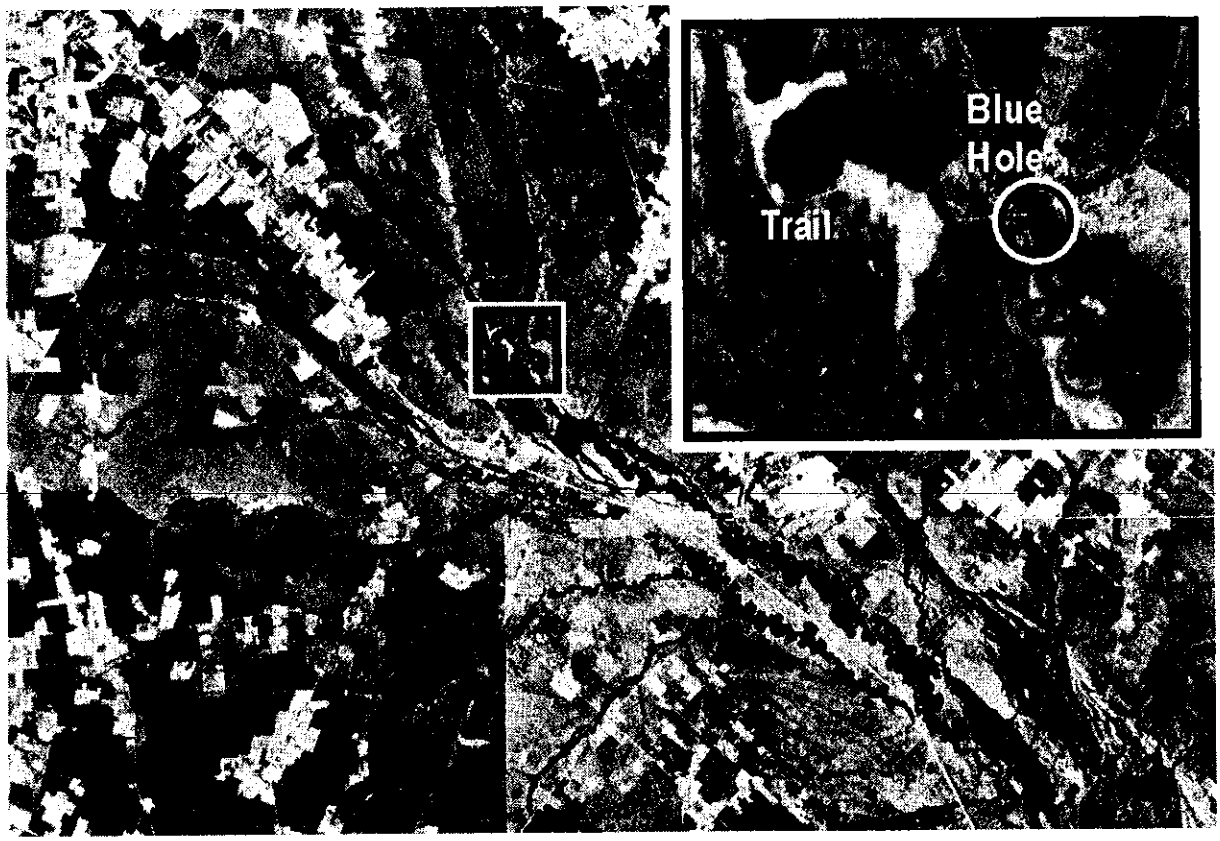


Figure 9.
In Gloucester County, Inskeeps Blue Hole is a site brimming with mystery, and replete with fanciful tales. Long touted as bottomless, it is the source of a deep spring on the Great Egg Harbor River. A hill capped with resistant gravels dams the river at this point, providing the Long-A-Coming Trail its natural causeway across the wide, wet paleovalley. It also would have been an ideal location for a springhead to have existed during the periglacial periods. Note the braided, high energy, high-flow stream bottom, characteristic of ephemeral waterways above the Inskeeps Blue Hole, and the perennial spring fed beaded drainageway below it. Both patterns are characteristic of the High Arctic today. Adapted by M. Mihalasky and T.S. Del Sontro from Airplane Atlas Sheets, 1930/32, No. 192, 193.

Below the Inskeeps Blue Hole, the river valley looks very different. A series of beaded, circular channels appear. In the Arctic, similar rivers with cuspated banks are attributed to a series of pools, created by a stream flowing along the lines of patterning associated with ice wedges (Everdingen, 2002; Hamelin and Cook, 1967, p.188). Such unusual beaded channels are unique to periglacial environments (Vandenberghe and Woo, 2002, p.495), and could be explained by the presence of a perennial spring when a cold climate was present. The spring would provide the water needed to maintain a year-round pool of unfrozen water. Coincidentally, the Lenape name for the site where the beaded Great Egg Harbor River and the beaded Pennypot Stream meet is descriptive of ponds. Penipach ghihil'len, meaning "it falls off" designates this confluence, (Stewart, 1932, p. 36) with the "pach" portion meaning "pond" (Stewart, 1970).

Another blue hole at Mount Misery, like the Inskeeps, issues from a broad valley bottom adjacent to a stream. Its appearance also demarcates the transition from a broad ephemeral paleochannel, to a beaded paleochannel along the North Branch of the Mount Misery Brook. Both the Mount Misery and Inskeeps springs occur where a waterway crosses a divide capped by sand, gravel, and clay. Curiously, the Mount Misery spring, like the Inskeeps, is also called "The Blue Hole" (Patterson; Gordon, Ted, pers. comm., 2003).

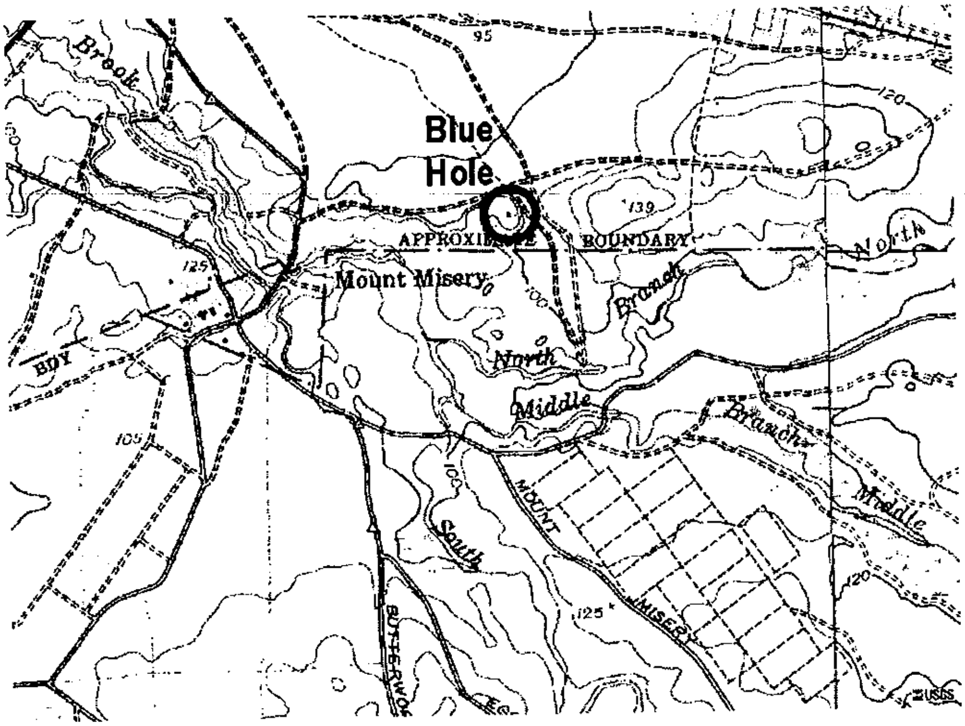
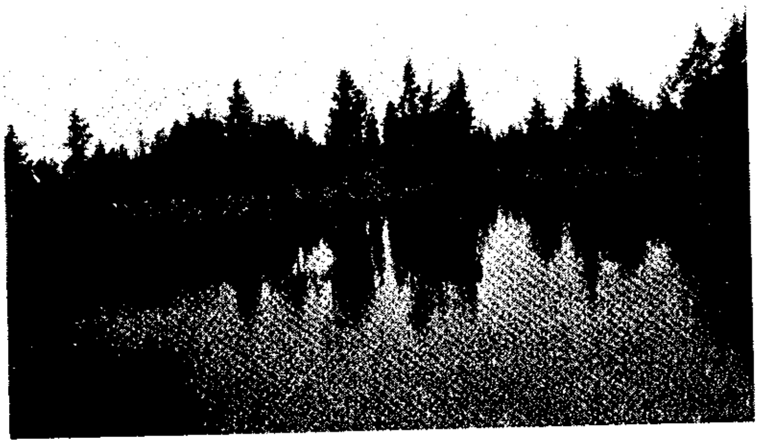


Figure 10
The Inskeeps Blue Hole has an obscure analog in Burlington County, also called the "Blue Hole". It too is below a natural dam, and demarcates the transition from a broad ephemeral paleovalley, to a perennial beaded drainage system. On the USGS Browns Mills Quadrangle, the tiniest blue dot marks the springs location. It is no accident that a time worn trail is adjacent to such a reliable water source.

Like spungs, blue holes often had precontact trails which were contiguous to their source. A ford of the Long-A-Coming Trail crosses the Great Egg Harbor River beside the Inskeeps Blue Hole (Boyer, 1962, p. 64), and is next to the site of long aboriginal occupation from the Archaic through the Woodland times (Mounier, 1972). A trail crossed Stony Brook at Dog Heaven, near a charcoal camp (Egg Harbor Township Tercentenary Publications Committee, 1964, p.68). The Davenport Hole (Mizpah) on the South River had six trails converge there, and served as a gathering site for local charcoalers.

Dick's Hole is another deep opening in a small branch of Cooper's Creek. It first appears on a 1686 map (Farr, 2002), and suffers the most unfortunate name for a blue hole the author has found yet! The old Egg Harbor Road passes alongside this spring on its way to the Inskeeps Blue Hole. Recent research by road historian Edward Fox is beginning to



correlate a series of ponds and springs in Camden County along this trail, from downtown Camden to the Blue Hole at Inskeeps Ford. This is the same trail as the Long-A-Coming which Chalmers chronicled (1951).

Figure 11
Savannahs once occupied several broad, flat paleovallyes of three separate watersheds in the Pinelands. Today they are basically limited to the Wading River area. A savannah's existence depended upon a combination of frequent fires, and a seasonally high water table of sufficient quantity and duration to inundate its meadow. The flooding discouraged tree reestablishment after a fire. Harshberger's photograph of a savannah along the North Branch of the Wading River illustrates that the meadow is scarcely above the water-line (1916, Figure 120).

OF SAVANNAHS AND PLAINS

In modern usage, savannahs are grasslands located in tropical and semitropical regions with a distinct wet and dry season (Burchfield, 1982). For example, in Africa, they are large, open plains of tall grassland, with a scattering of drought-resistant trees. In contrast, 16th century Spanish explorers, who first recorded the term, used it to denote a specific type of wetland. They described "savannahs" as the treeless, open and marshy plains in tropical America, chiefly in the Caribbean (*Oxford English Dictionary*, 1971). By the 18th century, the term was employed throughout the southern Atlantic coastal region for labeling low-lying, marshy meadows, often interspersed with pines (Hall, 2002). The early residents commonly used savannahs to pasture cattle from the tropics, north to New Jersey.

Savannahs once covered thousands of acres in the interior of Atlantic, Burlington and Ocean Counties (Harshberger, 1916, Plate I). Harshberger's savannahs are clearly marked as sparsely wooded pine swamp on the topographic sheets of Cook and Vermeule (c.1889). A scattered nature for the forest is inferred from the wide spacing of the pine tree symbols (i.e., the icon density).

Wet meadows could be found on the wide drainageways that flowed into the mid- and upper-reaches of the Great Egg Harbor, Mullica, and Wading Rivers. These meadowed drainageways are braided paleovalleys, related to nival flow during the cold period of the Late-Pleistocene (see "Of Cripples and Spungs, above). The braided stream deposits, which Farrell et al. (1985) reported at Atsion, are listed as savannah on Harshberger's map. Today, savannahs are limited to an area of no more than 1,000 acres (McCormick, 1979, p. 231).

Some confusion exists with the current usage of "savannah" in the Pinelands. A sparsely wooded upland is occasionally referred to as a savannah. The name suggests an open, park-like dry woods, perhaps akin to the plains of Africa. In New Jersey, a savannah has historically referred to a grassed wetland with a scattering of pines. Open tracts on uplands, with low pines and stunted growth, are referred to as "plains". Now limited to the

Pinelands core area (Good, Good and Andresen, 1979), extensive tracts of the scrublands once were scattered throughout the Pine Barrens. The name "Belleplain" designates the area where a large plain once existed (Horner, 1850). The Risley Plain in Weymouth Township became part of the Estelle Colony.

ORIGIN OF SPUNGS

In his inventory for the National Park Service, Jack McCormick (1970, p.82, 83) recognized the outstanding potential for geologic study the Pine Barrens offered. He states:

> Our knowledge of the geologic and biologic history of the region is poor and conflicts of opinion have arisen frequently between adherents of opposite views.... The physical environment of the Pine Region during the Pleistocene Epoch and during the subsequent warming period are virtually unknown. Some geologists claim to have found evidence of severe tundra-like conditions, but others claim the region only slightly cooler and wetter than the present. ... Another geologic mystery concerns the numerous saucer-shaped depressions scattered throughout the Pine Barrens. These depressions, similar in many respects to the better known Carolina Bays, have been attributed to phenomena of original deposition of the Coastal Plain formations, to differential illuviation of clay and silt particles and their deposition in the depression areas, to wind erosion, and frost-thaw action during the Pleistocene.

McCormick considered the Pinelands an advantageous place for researchers because of its proximity to academic and governmental institutions, its pristine state, and its simple terrain.

Three popular hypotheses have been proposed for the origin of spungs. The earliest, Wolfe's "frost thaw" and "thermokarst-lake-basin" hypotheses (Wolfe, 1953; 1977, p.290-293) are vaguely written (French and Demitroff, 2001, p.341), and can be combined as a single hypothesis for convenience. Archeologists Anthony Bonfiglio and Jack Cresson proposed a "pingo scar" hypothesis (Bonfiglio and Cresson, 1982) for spung formation at the Third Annual Pine Barrens Research Conference, held in 1982 at Stockton State College, in Pomona. The "periglacial wind action" hypothesis by French and Demitroff (2001) is the most recent one to explain the origin of the numerous wetland-filled depressions in the Pinelands.

In August, 2000, French and Demitroff began a series of field investigations to search for evidence of southern New Jersey's Pleistocene inheritance. From this research (French and Demitroff, 2001; French, Demitroff and Forman, 2003), one can infer that conditions would have been cold, windy and arid, with episodes of permafrost for at least two periods of the Late-Pleistocene. In all likelihood, the vegetation coverage would have been sparse, with significant eolian activity.

A DWINDLING NATURAL AND HISTORICAL RESOURCE

Many elderly residents of the Pinelands tell stories of how the water has disappeared from cripples, swamps, streams and spungs over their lifetime. The author has interviewed many an old timer in the pursuit of preserving the local lore. This recurrent theme, if true, has many implications for the overall well-being of the area's flora and fauna.

Accounts of the region's desiccation go all the way back to the time of Peter Kalm, the botanist. Upon his visit to Raccoon (Swedesboro) in November 1748, he records some accounts of smaller lakes, brooks, springs and rivers drying up over a period of many years (Kalm, 1770, p. 285-187). The flow of larger rivers was reported to have remained about the same. William Farr, in his *Waterways of Camden County* , comments that "almost all streams have suffered a diminution in flow and size" in the region, and "many small streams ... have disappeared" altogether (p. ii-iii). Accompanying the drought of 2000 to 2002, a number of newspaper and magazine articles addressed the apparent problem of South Jersey's dwindling hydrologic resources (e.g., Kaskey, 2000; Hajna, 2001; Moore, 2002). What is not readily determined is the cause, and who is responsible for monitoring shallow groundwater levels. Stockton College has begun to address this phenomenon (Del Sontro, 2003; Mihalasky, this volume). The Pinelands Commission does not view the issue of the ponds drying up as a problem (Statewide Watershed Management Advisory Subcommittee, Meeting Highlights, December 12, 2000).

Spungs as Indicators of Hydrologic Change

An important reason why South Jersey's spungs have been overlooked is that they remain waterless for longer and longer periods, and many have dried completely. Three large ponds, linked by the Egg Harbor Trail, are used as examples to illustrate this point. Each of these watering holes was once a well-recognized landmark by 17th, 18th and early 19th century travelers. A Colonial traveler from the Delaware River would use this trail network on his way to Egg Harbor, then the name for eastern Atlantic County, and possibly also the name of a village now lost to history (Mickle, 1845, p.111, 112).

We begin with the Broad Pond, a mega-spung that once was the largest water body in Salem County. It is positioned between Elmer and Willow Grove, in the Pine Barrens section of Pittsgrove Township. The author learned of this site when hired to evaluate some trees for the Broad Pond Refuge, a private animal preserve. The owners assumed that Broad Pond was the small excavated fishpond next to their residence.

Upon research for the provenance of its name, it came to light that Broad Pond was a prominent feature on many old maps and surveys (Nelson,1781; Everts and Stewart, 1876; Woodward, 1964, p. 8,9). The Egg Harbor Road crossed the southeast corner of this spung, on its way through the woods towards Vineland. The region southeast of Broad Pond was called "Broad Neck", a desolate pine barrens whose "inhabitants were not noted for their enterprise, or for being very unexceptional citizens" (Cushing and Sheppard, 1883). Broad Pond is, today, a low-lying woods, and has disappeared even from the memories of Pittsgrove Township's oldest residents.

Continuing eastward 8 miles by way of the Forked Bridge (Willow Grove) over the Maurice River, it is on to the next major spung named "Egg Harbor Pond", at Marcacci's Meat Market in Vineland. The Hollingshead Survey (1795) of 19,623 acres, drawn from the Penn's survey, clearly marks the location of the pond along the Maul's Bridge section of the Egg Harbor Road. From the ground, there is nothing to indicate that this once contained a great body of water. However, from an oblique air photo taken during the wet winter of 2001, its parabolic outline is easily discerned in the open farm field (French and Demitroff, 2001, Fig. 2A). Many spungs probably share Egg Harbor Pond's rounded saucer-like shape, except their margins are often obscured by thick vegetation.

John Marcacci, now deceased, recounted in an interview (2001) with the author his childhood memories of Egg Harbor Pond, when water still filled it year-round. He stated that, in the late 1920s, the pond had sufficient quantities of water to support fishing, boating and swimming. In the spring, its margins would almost reach Vine Road, several yards higher than the ancient pond bottom. Vine Road had to end its westward trek at the Central Jersey railroad track, since the terrain was too wet to continue on to Main Road. Now Vine Road continues to Main Road over dry ground. Powerful springs fed Egg Harbor Pond, which the Vineland Railway tried to cap, unsuccessfully, by dynamiting them around 1870. By 1960, the pond had lost so much water that a slough was dug for irrigation purposes in its center, as an attempt to reach the dropping watertable.

The third prominent spung is 6 miles away on the Egg Harbor Trail, called Horse Break Pond in Buena. On a survey (Wright, 1867), the Egg Harbor road again strikes the southeastern corner of a spung. A surveyor created the map for the bitter land battle against the much reviled Richard West of Catawba fame, a man imprisoned for forging deeds. His story is one of suspected murder, mystery and greed in the Pine Barrens and has been fodder for much scandalous lore (Boucher, 1963, p. 50-51; Barrett and Scull, 1968). The site is purported to be adjacent to an early Swedish settlement (Reed, pers. comm., 2000), complete with a burying ground. A bouncer at the Midway Tavern removed the gravestones for use in building a walk during the 1940s.

In the mid-1800's, a charcoal camp known as "Stephan Colwell's Coal Grounds" (Beers, 1872b) developed adjacent to the pond. It is the author's belief that there is a significant relationship between the charcoal camp and the Underground Railway. The nearby Mt. Union AME Church, a log structure, was built in 1858 as a sister congregation to the Springtown AME in Greenwich (Trusty, 1997). The black abolitionist William Still was a prominent patron who financed escape networks throughout the Delaware Valley. He made his fortune through the charcoal trade in Philadelphia. As a young man, he and his brother James Still, the famed "Black Doctor of the Pines", cut cordwood for charcoal at the Brotherton Indian Reservation (Khan, 1872; Still, 1877). The Still family remains well represented in the area today.

Like the Swedish settlement and charcoal camp, Horse Break Pond is likewise forgotten. The meadow which marks the spot has a foot or so of water every now and then. But this is not quantity enough, nor is it filled often enough, to prevent tree encroachment. The pond is not expiring from infill, but is simply starved of groundwater.

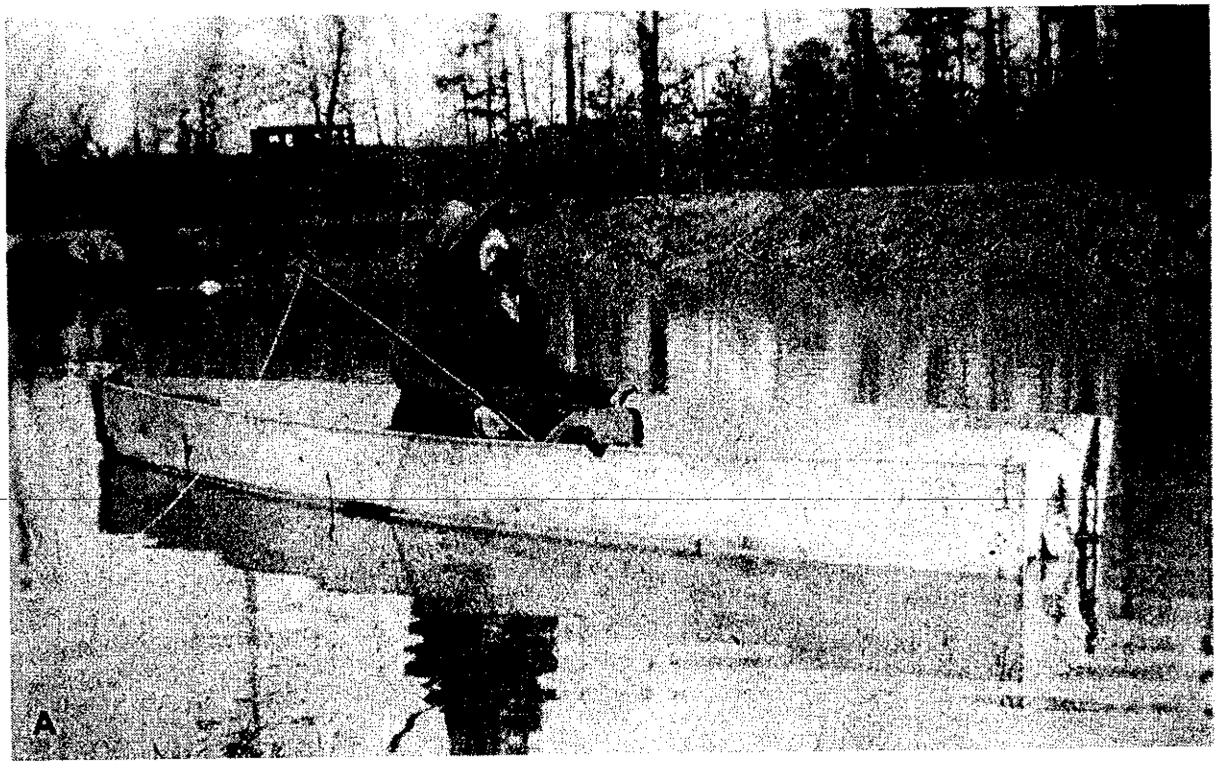


Figure 12
The Punch Bowl was a circular spung a mile north of Milmay on Tuckahoe Road, fronting on the Hensel Farm. Water filled it for much of the year from the 1890's to the 1930's. A prank was captured on this photograph from ~1930 (A) taken by John Madera, Sr., and distinctly shows the former level to which this spung would normally fill. The effigy is a caricature of Old Man Hensel in Madara's boat, "gone fishing". Although the Punch Bowl was barren of fish, other nearby spungs had sizable fish populations (Madara, who provided the photograph, pers. comm., 2003; Hensel, pers. comm., 2003; Fanucci, pers. comm., 2000). Today, what remains of the Punch Bowl (B) never contains much water at all, even in the wettest of seasons. It is widely accepted that the region's watertable has dropped significantly in the last 50 years, according to Madara, Hensel and Fanucci.

Cripples as Indicators of Hydrologic Change

Cripples can also provide us with clues as to the health of an area's groundwater. As stated earlier in the section on shallow hydrology, runoff is negligible in the Pine Barrens. A cripple depends upon shallow groundwater for recharge, as do Pine Barrens rivers and spungs. Bob Francois, historian, has long-time family connections to the Union House above Millville. Settlers arrived at the site prior to 1729 on the Cape Road, where it crossed the Maurice River. A nearby cripple had an interesting metamorphosis of name. Francois's grandfather, Albert Bebee told him of how the Cape Road forded a small, perpetually wet valley just to the east, and so it was called "Horse Wallow". A map of Cumberland Co, New Jersey distinctly indicates water in Horse Wallow (Beers, et al., 1862). Over the years, the spot became dryer and dryer. By 1890 the name of the crossing became known as "Horse Hollow", since it was no longer wet and remains dry today (Francois, pers. comm., 2003).

Springs as Indicators of Hydrologic Change

Many springs have disappeared over the last century. Pamphylia Spring, on the banks of the Cohansey River in Bridgeton, was once so mighty that in 1716 Pamphylia Township was proposed in its honor (Heston, 1924, p. 821; Elmer, 1869, p.30, 48). Jim Holder, local archeologist, is the only individual I have found who remembers this spring. During the 1930s, fisherman could still fill their jugs with its health-giving waters to slake their thirst during a hard day's outing. It had a modest flow in the 1930s, and the high tide would cover the source. Still, it would bubble above the brackish water (Holder, pers. comm., 2002). Its whereabouts are not apparent today.

The "boiling springs" of the Manumuskin River had wide recognition in the 19th century when Bowen wrote his *History of Port Elizabeth*. The sparkling and hygienic water bubbled profusely from the river banks (1885, p.39). Despite the relatively pristine state of the river's surroundings today, the springs are gone.

When Kathryn Chalmers (1951, p.84-86) played on the banks of the Great Egg Harbor River as a child of the 1880's, she recounted the Inskeeps Blue Hole as a fantastic feature with abundant springs. Its surcharged flow would create a whirlpooling effect, adding to its mystery. Beck reported in the 1930's that the hole was still sinister, but was probably larger some years ago (1937, 148-153). Today, it is a shallow pool, incompletely filled, and without a hint of its former blue tint. Like the Dog Heaven in Farmington (Egg Harbor Township Tercentenary Publications Committee, 1964, p.68), and the Danger Hole by the author's home along the South River, and many other nameless blue holes throughout the Pines, its artesian head has long-since faded.

Savannahs as Indicators of Hydrologic Change

Silas Little theorized that wetland meadows could have been formed in the Pine Barrens as a result of dry season wildfires (1946, p. 37, 39). If a forest fire consumed the wetland's trees and shrubs, but left the organic mat unchanged, woody plant regrowth would have occurred rapidly. However, if the organic layer was consumed to the watertable, then the woody vegetation's regrowth would be delayed until the *Sphagnum* mosses accumulated in sufficient quantity to create a new seedbed (Little, 1979, pp. 308-310).

Whittaker agrees that dry season fires, and soil factors, probably created the savannah-like communities which Harshberger described (1979, p.320). A high watertable is necessary to have savannah-like conditions. At the braided stream deposits near Atsion described in Farrell et al. (1985), a severe fire in 1977 burned the covering vegetation and the organic mat down to the mineral soil. Eighteen months after the fire, pine seedlings sprouted everywhere, except in the lowest troughs. Today, the Pitch Pine forest reaches 30 feet in height at the research site. Meadow-like conditions are limited to the interstice formed by the lowest channels between the ancient bar deposits. (Farrell, pers. comm, 2003). Harshberger (1916, Plate I), and Cook and Vermeule (c.1889, sheet 15), indicate a savannah once existed at this spot.

In Little's (1979) model of meadow formation, trees should not have reestablished themselves at the Atsion site until a layer of *Sphagnum* mosses accumulated. Instead, the area quickly reforested, despite an insufficient amount of time for the mosses to have amassed. The mineral soil remained dry enough to promote the germination of woody plants. Perhaps the disappearance of savannah vegetation at the Farrell site, as well as at other savannahs sites throughout southern New Jersey, could be explained by a recent lowering of the water table.

CONCLUSION

During the Late-Pleistocene, southern New Jersey was at times a cold, arid, and windy plateau. The geologic evidence suggests that climatic forces much different from those of the present reshaped the Pine Barren's landscape. Periglacial processes provided the conditions needed to create some of the peculiar features found in the Pine Barrens today, including spungs, cripples, blue holes, and the paleovalleys which once supported savannahs.

The water table rebounded to near present levels some 12,000 years ago, as a result of the moderation of climate and the associated sea level rise. A higher water table provided a ready supply of water, and abundant life returned to the Pinelands. For the next 12,000 years, various populations, both human and beast, utilized the spungs and blue holes inherited from the region's periglacial past.

Spungs are literally windows to the ground watertable. Their replenishment depends almost entirely upon seepage of groundwater from the uppermost confining layer. By observing how much water a spung holds, and for how long it holds it, insights can be gained as to the region's hydrologic soundness. Using a combination of historical records and anecdotal comments, it is possible to draw a reasonable conclusion as to the general trend of groundwater levels over time.

The author concludes that the watertable throughout South Jersey has been slowly declining for the past 250 years. This loss of groundwater has accelerated since the 1930's. Large, perennial ponds, where residents once fished, boated and swam, dotted the Pinelands a century ago. Little evidence remains that they ever had been ponds. Cripples that in the past mired wagons, are today dry hollows. Several centuries ago, strong springs flowed with such great force, they inspired tales of whirlpools, bottomless chasms, and sparkling blue waters. They are gone too. So are thousands of acres of savannah wetlands.

Spungs are not irrelevant, nor are they expendable. They possess significant potential for future historical and scientific research. Spungs serve as important habitat for rare and endangered flora and fauna. For certain, we know that the demands placed upon South Jersey's water supplies are increasing. What's unknown is how over-withdrawl from deeper aquifers has affected, and will affect, the fragile ecosystem so dependent upon a ready supply of near-surface water. The shallow aquifers are the lifeblood of the Pinelands.

ACKNOWLEDGEMENTS

I would like to thank the following for the help and support: Hugh French, who has most generously provided his time and expertise in investigating southern New Jersey's periglacial geomorphology, and furnished constructive comments in the preparation of this paper; Mark Mihalasky, for his assistance in the field and in development of this paper; Michael Hozik, who motivated the author to push forward with his when inspiration for periglacial research was short in supply; Julie and Fred Akers, of the Great Egg Harbor Watershed Association, for their keen interest and support; Paul Schopp, Ed Fox, and the other fellows of the West Jersey Historical Roundtable for their insightful commentary on southern New Jersey's past; Joe Arsenault, for sharing his knowledge on the region's wetlands; Harry and Gail Benson, for their technical support; Lois Mazza and the Reference Department at the Atlantic County Library, for their assistance; Christian Bethmann for locating the Mt. Misery Blue Hole; Ronny Bertonazzi, Glen Dalponte, Richard Hanson, and all the many local history devotees who share a passion for the outstanding lore and natural beauty of the Pines; and last, but not least, my wife Patricia, who tirelessly and patiently assisted me in the field, with the preparation of graphics, and in coordinating all the multifarious duties which made this paper feasible.

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Is the Pine Barrens Water Table Declining? What Does the Record Show?

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ABSTRACT

There is much concern about water table decline in the New Jersey Pine Barrens and its impact on wetlands and streams. The U.S. Geological Survey has been monitoring water tables since 1936 but the number of wells, the frequency of their reading, and the length of their records all vary, resulting in a spotty database. Regardless of these weaknesses, some trends are evident but they are not uniform. Seasonal high water table at some sites has risen, at other sites has fallen, while at still others sites has stayed about the same over comparable periods of record.

INTRODUCTION

Concern that the Pine Barrens water table is declining comes up when wetlands appear to dry up and streams cease flowing. For example, French and Demitroff (2001) partially attribute the origin of enclosed Pine Barrens depressions to water table decline. The purpose of this study is to see what records exist that document changes in the region's water table. The source of this information is the U.S. Geological Survey's historical ground water information available on their web site (wwwnj.er.usgs.gov).

The Nature of the Record

Until recently, the U.S. Geological Survey's web site had a special button for water table aquifers that made the search for information for this study simpler. (Currently, this information still exists but is incorporated with data for confined aquifers.)

There are 14 wells monitoring the water table in the Pine Barrens (i.e., the Kirkwood-Cohansey aquifer; Figure 1, Table 1). These wells are distributed throughout the Pine Barrens. Also the depth to the water table is also varies. Some are deep while others are shallow. But there are "holes" in this database.

First, the length of record of these wells varies from a maximum of 65 years for the well at Penn State Forest to a minimum of 12 years for the well at the Washington Township Municipal Utility Authority (MUA).

Second, and more of a problem, is that the U.S. Geological Survey took episodic readings that varied in annual frequency. Ideally, one reading per month would adequately cover the annual water table fluctuations. But in some years and at some localities, far fewer readings were taken and in some cases even suspended (Table 1).

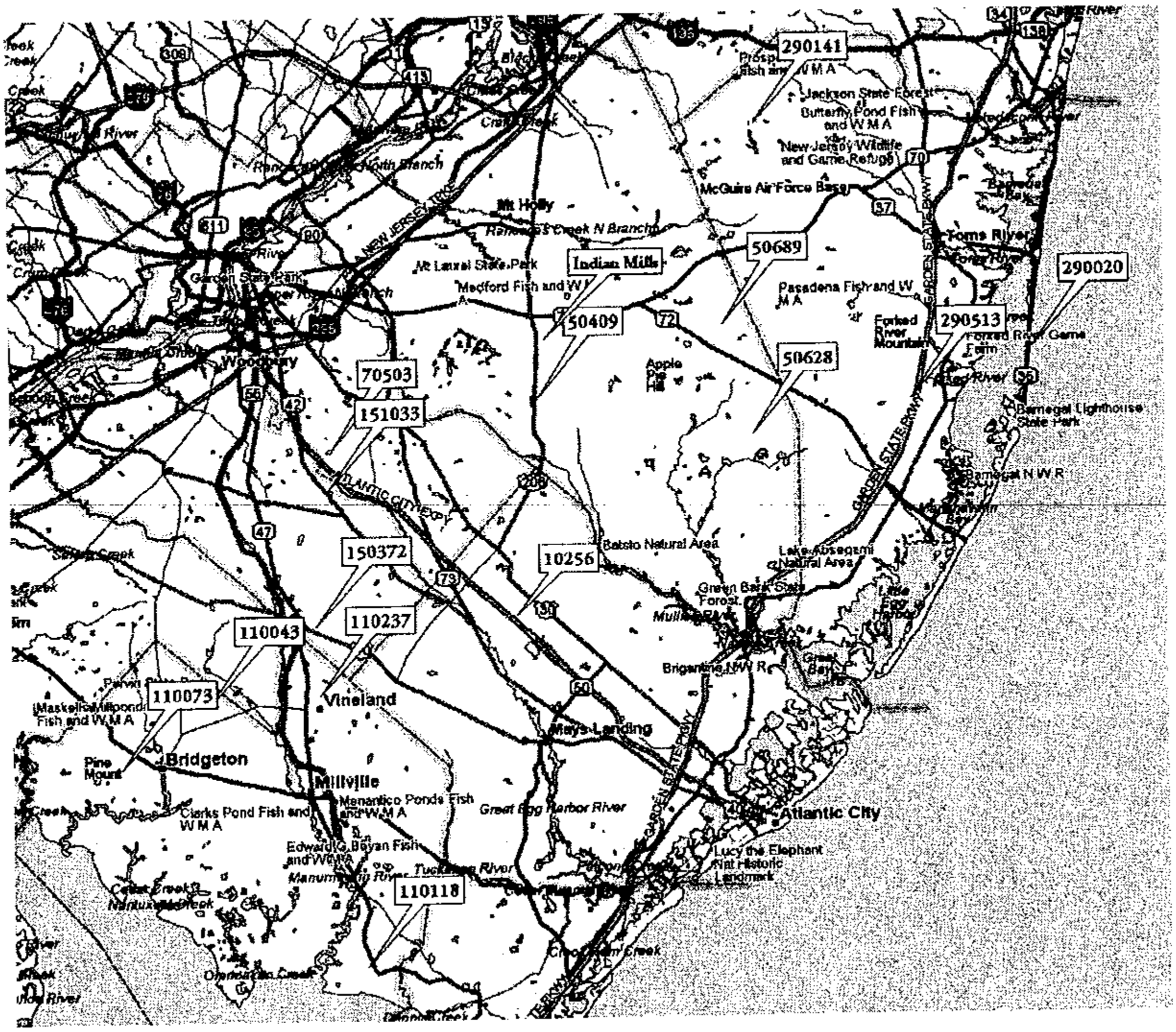


Figure 1. Locations of wells considered in this study.

METHODS

Seasonal high water table marks the hydroperiod when ground water is either closest to or even at or above the ground surface. This is the time when wetlands and streams receive their maximum ground water discharge. The shallowest depth to water was taken annually from each site for its entire period of record. This is referred to as the seasonal high water table. (The end of the period of record was the year 2001.)

Next a five-year moving average was calculated for each locality. The first calculated value occurred after the first five years of annual data accumulation. In one case, Newfield Water Department information was absent for three years and so the five-year average began five years after the gap ended.

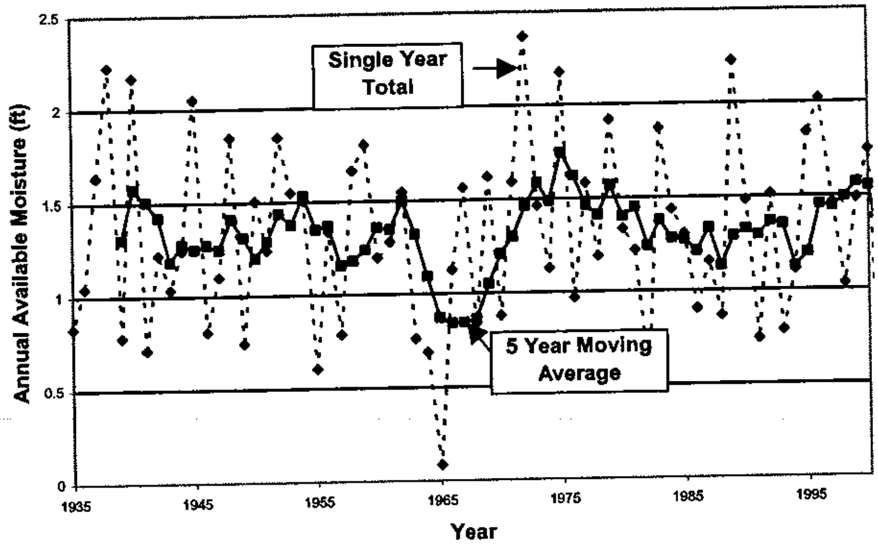
Table 1. Water Table Wells in the Kirkwood-Cohansey Aquifer System

Well Number	County	Well Name	Average Water Table Depth(ft)	Total Years	No. Years With <4 Readings Per Year	Percent Record Based on <4
10256	Atlantic	Scholler Inc.	36.15	40	8	20.0
50628	Burlington	Penn St. Forest	1.52	65	9	13.8
50689	Burlington	Lebanon St. Forest	19.77	46	3	6.5
50409	Burlington	Atsion	5.56	38	14	36.8
70503	Camden	Winslow WC	31.57	29	6	20.7
110118	Cumberland	Heislerville	3.19	30	30	100
110237	Cumberland	Vineland	8.38	30	30	100
110043	Cumberland	Deerfield Cum.Co.Tec	4.90	30	30	100
110073	Cumberland	Holly Shores	4.68	29	29	100
150372	Gloucester	Newfield WD	19.03	15	4	26.7
151033	Gloucester	Washington Twp. MUA	15.14	12	0	0
290513	Ocean	GS Pway.(Rt.532)	5.72	40	29	72.5
290020	Ocean	Long Beach Island	3.52	40	22	55.0
290141	Ocean	Colliers Mills	4.50	25	0	0

Finally an annual climatic index was formulated to determine how much rainfall and evapotranspiration occurred annually. This was accomplished by talking the monthly total rainfall and monthly average temperature for the weather station at Indian Mills, New Jersey. This station was chosen because of its length of record extended further back than the longest monitoring well period of record, and because of its proximity to wells in the most protected parts of the Pine Barrens. Thornthwaite's potential evapotranspiration calculation, as outlined in Gray (1970), was determined for the Indian Mills weather station for each month from 1936 to 2001. Next, the value of potential evapotranspiration was subtracted from the monthly precipitation. The resultant difference was called *available moisture* because evaporation and transpiration had been removed from the hydrologic cycle, leaving only surface runoff and groundwater recharge. Since surface runoff is thought negligible in the undeveloped Pine Barrens itself, *available moisture* is a reflection of how much water was available to recharge the water table (Chart 1). A word of caution, *available moisture* is an <u>index</u> of annual potential recharge used to distinguish dry years from wet years and is not meant to be an actual measurement of ground water recharge.

Next, the year's *annual available moisture*, from March through the following April, was summed. This interval was chosen because April is generally the time of highest seasonal water table. The *annual available moisture* was then compared to the seasonal high water table to discern whether difference in *annual available moisture* was responsible for water table fluctuations (Charts 2 through 13).

Chart 1. Available Moisture at Indian Mills, NJ



RESULTS

Twelve of the fourteen monitoring well localities were analyzed (Charts 2 through 13). The remaining two, Newfield and Washington Township MUA, have too short a period of record in comparison to the other monitoring wells.

Chart 2. Penn State Forest

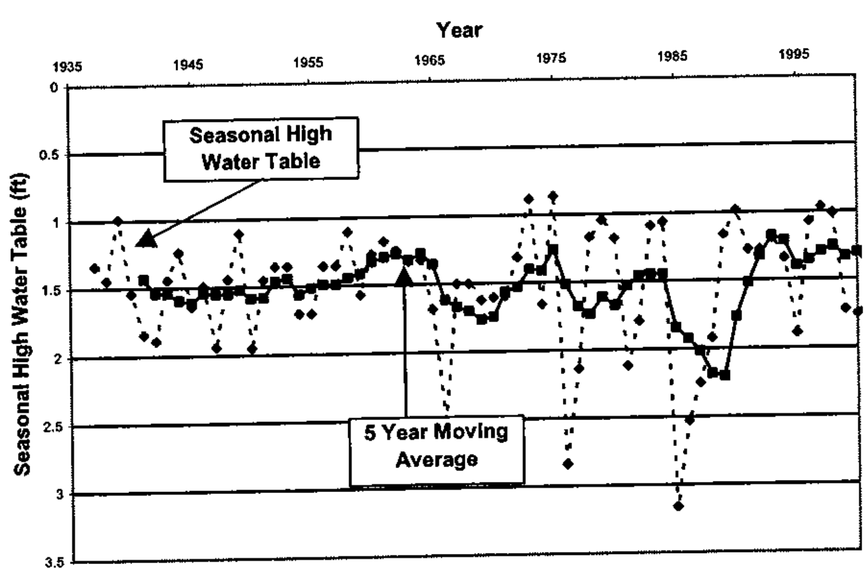


Chart 3. Lebanon State Forest

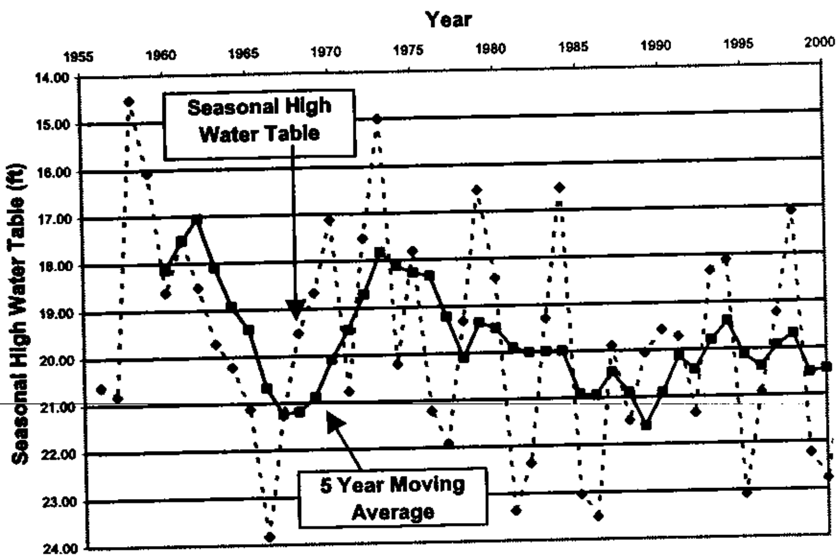


Chart 4. Scholler Inc. (Atlantic County)

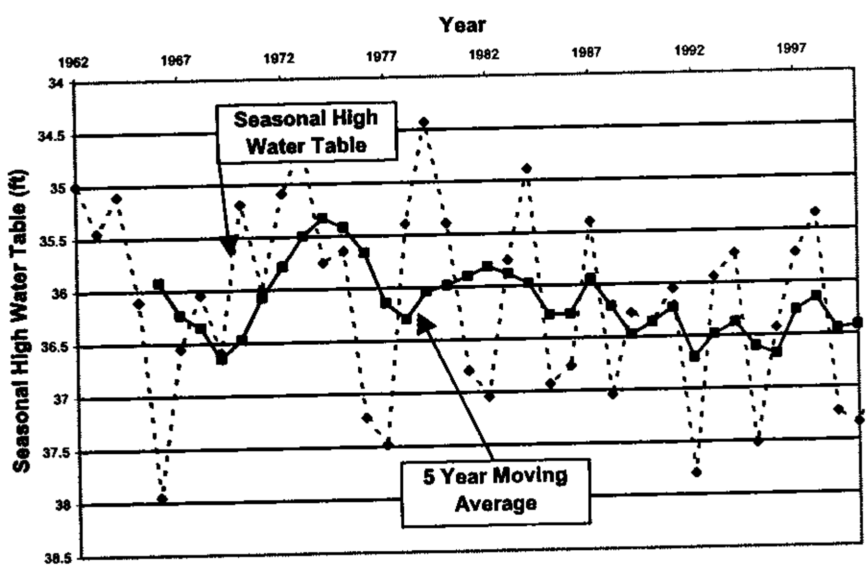


Chart 5. Atsion, NJ

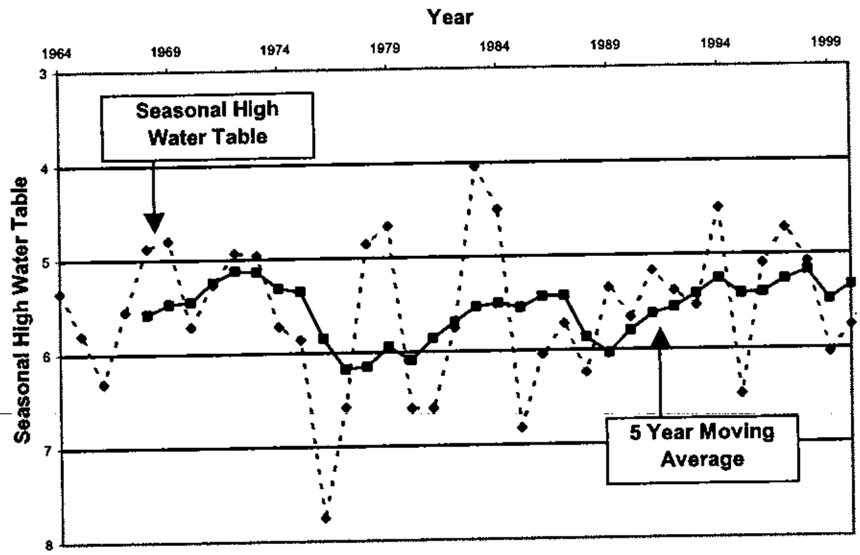


Chart 6. Garden State Parkway by Route 532

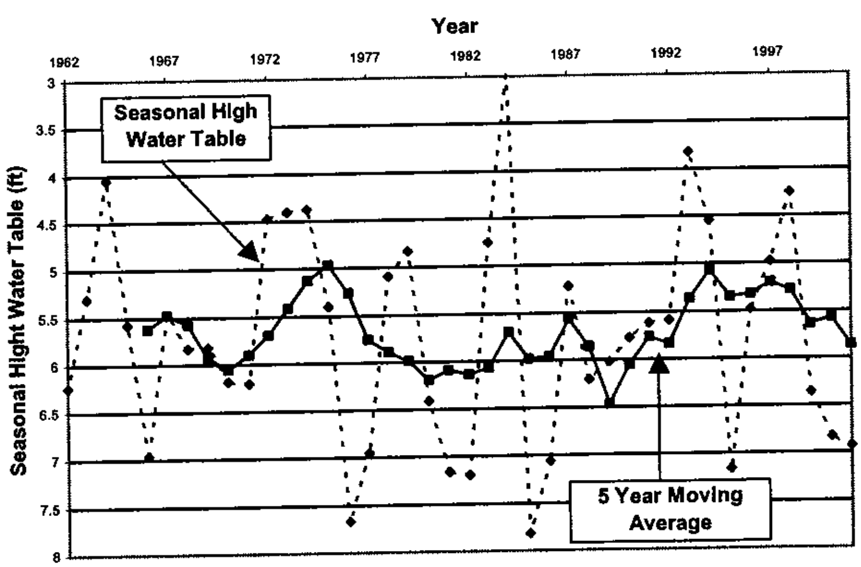


Chart 7. Long Beach Island

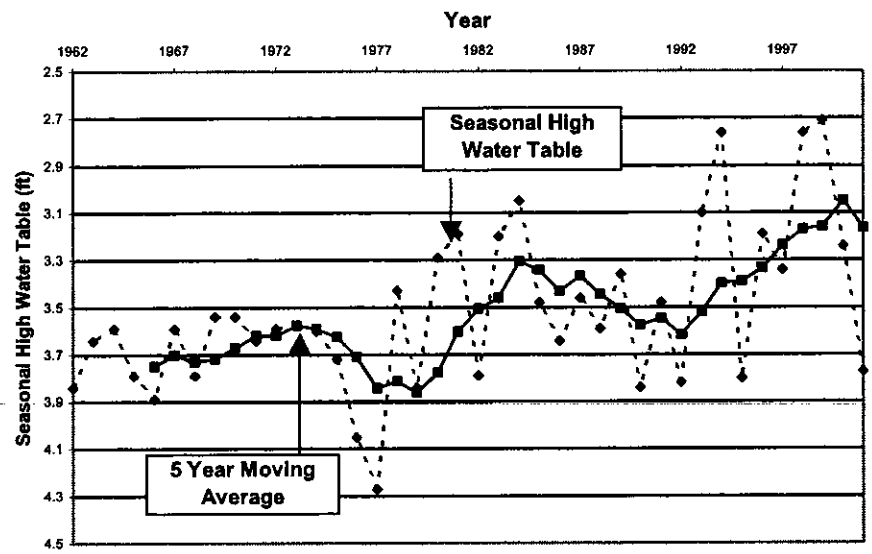


Chart 8. Heislerville, NJ

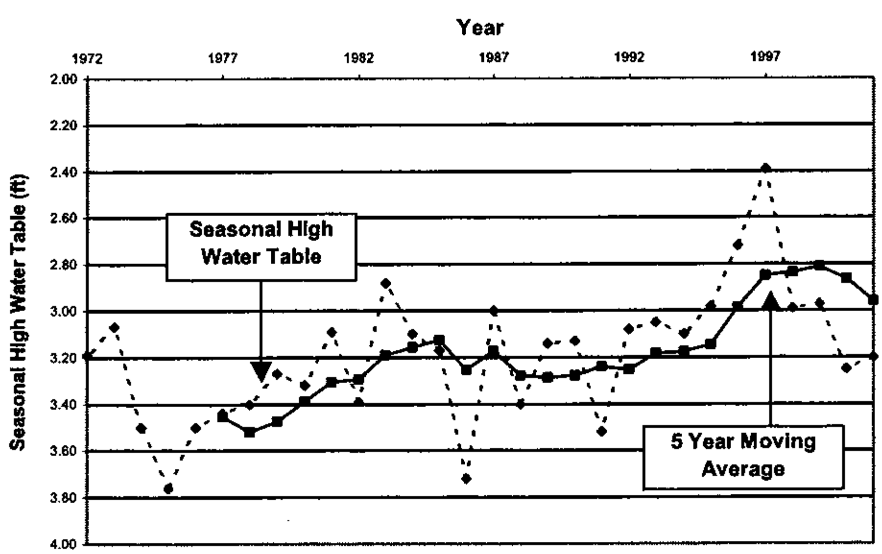


Chart 9. Vineland, NJ

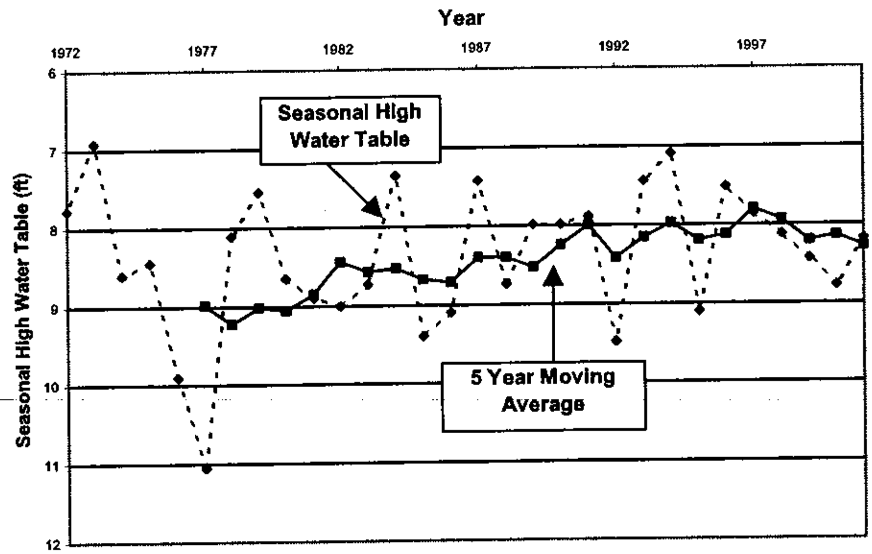


Chart 10. Deerfield Cumberland County TEC

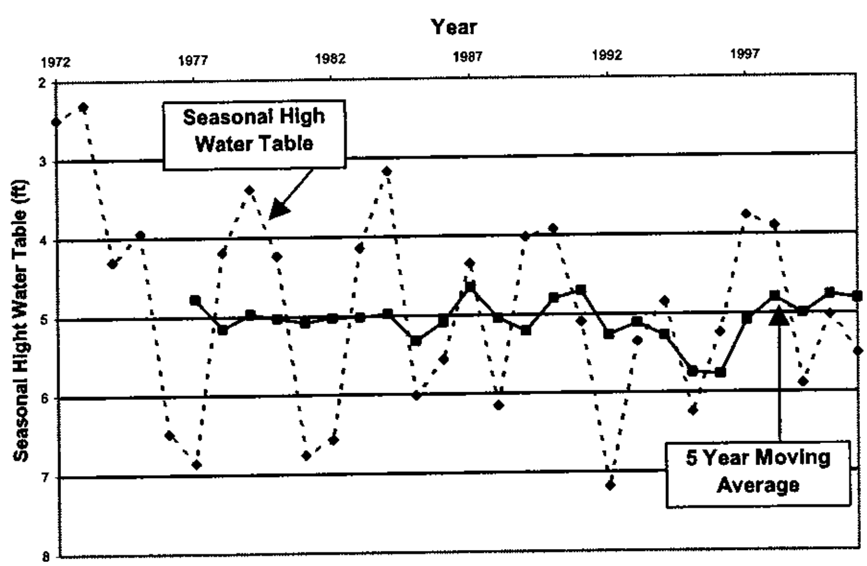


Chart 11. Holly Shores

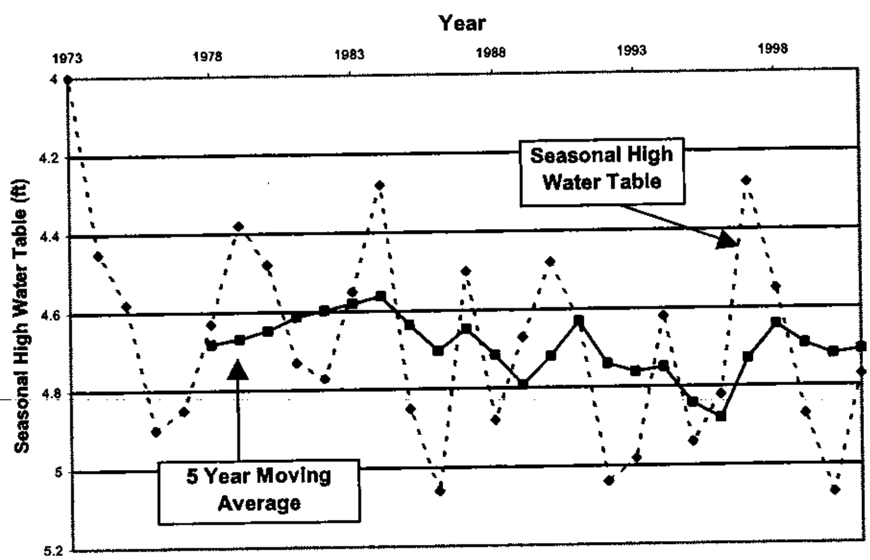


Chart 12. Collier's Mills (Ocean County)

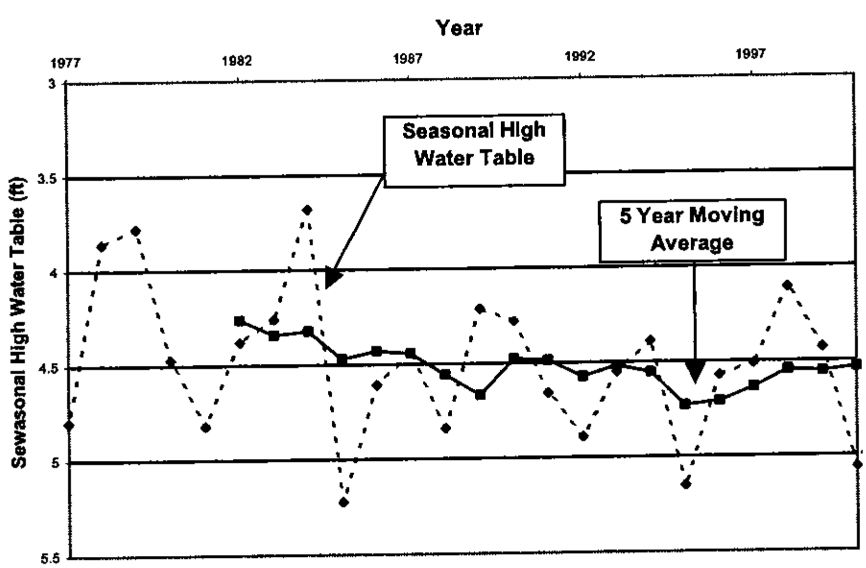
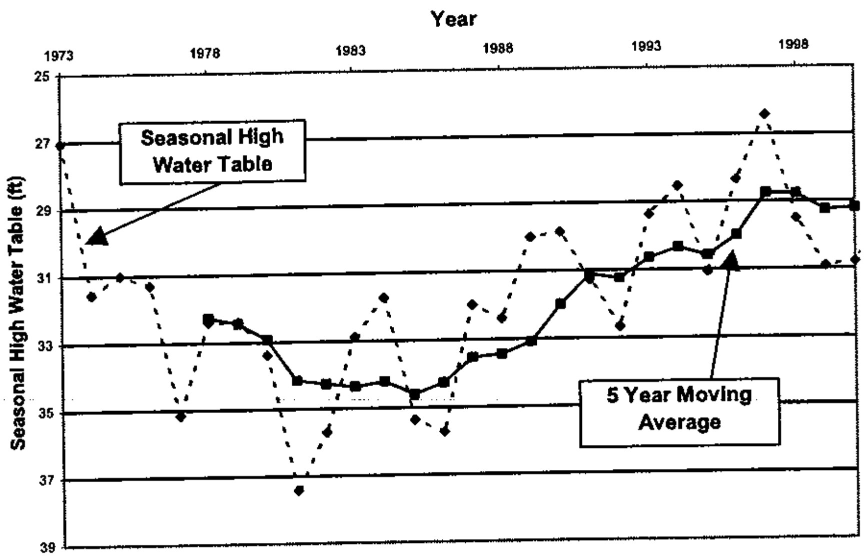


Chart 13. Winslow Water Company



A review of all of the hydrographs shows no regional pattern (Table 2). (The five-year moving average patterns are used in this discussion to avoid all the variation present in the annual patterns.) Some wells rise over the period of record while others fall. Still others are more of less steady.

Table 2. Seasonal High Water Table Change of the Period of Record

Locality	Change	Amount(ft)
Lebanon St. Forest	Deeper	3.5
Scholler Inc	Deeper	0.5
GSPway Rt(532)	Deeper	0.3
Collier's Mills	Deeper	0.2
Penn St. Forest	Shallower	0.2
Atsion	Shallower	0.2
Long Beach Is.	Shallower	0.6
Heislerville	Shallower	0.4
Vineland	Shallower	1.0
Winslow	Shallower	3.0
Deerfield	Steady	0
Holly Shores	Steady	0

Most of the changes are fairly subtle (i.e., less than 0.6 feet over the period of record). The water table decline at Lebanon State Forest and the rise of the water table at Winslow may be significant because of the magnitude of the change and the quality of their periods of record. (What both localities have in common is that the seasonal high water table in fairly deep for the Pine Barrens, 20 and 30 feet respectively.)

DISCUSSION

Regional Changes in Seasonal High Water Table

Twelve monitoring wells cover the same time period from 1973 to 2001, which affords the chance to see whether they have responded similarly over that time period. Each station was correlated with each of the other stations and the results are presented in Table 3. All the correlations were positive indicating they all tended to rise and fall at the same time. However, the correlation for many of these stations is weak to fair showing some possible similarity. (Again the annual frequency and timing of these readings differed from station to station and year to year, affecting the quality of the correlations.) Long Beach Island, Heislerville, and Winslow, in bold font in Table 3, stand out as not correlating with any other station (i.e., <.6000). Perhaps Heislerville's and Long Beach Island's coastal position is the cause for these. Winslow is another matter, being in almost the opposite hydrogeologic position (i.e., near the western margin of the outcrop belt), but the source of this poor correlation is far less certain.

Table 3. Correlation of Annual Season High Water Change Between Stations

	Penn	Lebanon	Scholler	GSP	LBI	Atsion	Heisler.	Vinel'd	Deerf'ld	Winsl.	Holly Sh.	Colliers
Penn	—	0.726	0.578	0.721	**0.296**	0.747	**0.301**	0.615	0.649	**0.584**	**0.544**	0.631
Lebanon			0.879	0.811	**0.227**	0.607	**0.132**	0.679	0.826	**0.553**	0.785	0.777
Scholler				0.757	**0.305**	**0.600**	**0.106**	0.730	0.892	**0.341**	0.841	0.799
GSP					**0.444**	0.764	**0.288**	0.704	0.699	**0.512**	0.605	0.700
LBI						0.409	**0.458**	**0.541**	**0.280**	**0.283**	**0.122**	**0.256**
Atsion							**0.401**	0.681	0.622	**0.460**	**0.500**	0.682
Heislerville								**0.374**	**0.174**	**0.487**	**0.236**	**0.036**
Vineland									0.722	0.636	**0.580**	**0.570**
Deerfeild										**0.533**	0.833	0.749
Winslow											**0.375**	**0.207**
Holly Sh.												0.701
Colliers Mill												—

Relationship of Seasonal High Water Table to Annual Available Moisture

The relatively low correlation coefficients could be the result of the method of "indexing" relative annual wetness or dryness as it incorporates many assumptions. First, the weather data consisted of a monthly precipitation total and a monthly temperature average from a single point (i.e., Indian Mills, NJ) in the Pine Barrens as compared to 14 monitoring wells spread throughout the Pine Barrens. Second, the time of seasonal high water table varies from year to year and sometimes from place to place. (For example, shallow water tables rise to the surface faster than deeper water tables.) One cannot be sure that the episodic readings, and the variable reading frequencies, taken by the U.S. Geological Survey actually "caught" the seasonal high. Third, the kind and intensity of its precipitation is ignored by the index. (For example, a year with many winter snowstorms probably results in far more ground water recharge than a year with many summer thunderstorms but both years might have the same annual precipitation.) Fourth, the Thornthwaite evapotranspiration calculation uses only two weather parameters, average monthly temperature and length of daylight. These are more aspects of climate that effect evapotranspiration than just these two (e.g., vegetation type, wind, humidity).

In spite of these sources of inaccuracy, all the correlations of annual available moisture and annual seasonal high water table were negative which seems reasonable. The less available moisture, the deeper the seasonal high water table (Table 4). (The use of the five-year moving average for both available moisture and seasonal high water table shows something similar but has two very weak positive correlation exceptions.)

Table 4. Correlation Between Available Moisture at Indian Mills, NJ and Seasonal High Water Table at the Monitoring Well Stations.

Location	Annual	5 year
Penn State Forest	-0.2076	-0.2665
Lebanon State Forest	-0.2237	-0.4768
Scholler Inc.	-0.0765	-0.3802
GSP(at Rt.532)	-0.0981	-0.2018
Long Beach Island	-0.0869	-0.1488
Atsion	-0.1888	-0.0175
Heislerville	-0.2184	-0.1883
Vineland	-0.0666	0.1425
Deerfield	-0.3918	-0.2052
Winslow	-0.1451	-0.3308
Holly Shores	-0.1529	-0.0868
Collier's Mills	-0.0758	0.1436

CONCLUSIONS

The U.S. Geological Survey maintains fourteen monitoring wells in the water table aquifer of the New Jersey Pine Barrens (i.e., Kirkwood-Cohansey aquifer). But the records of these wells have weaknesses. The periods of record vary but a more significant weakness is the variation in annual reading frequency from well to well and over time.

The direction of water table movement varies locally over the period of record. Seasonal high water table has gotten deeper at four localities. It has gotten shallower at six other localities. It has remained steady at two other localities and has too short a period of record to know for sure what the trend is at the remaining two localities. Most of the changes have been fairly subtle but two show significant changes. The water table at Lebanon State Forest has gotten 3.5 feet deeper while that at the Winslow Water Company has gotten 3.0 feet shallower.

The annual regional change in seasonal high water table shows weak correlations between most wells. This weakness probably reflects the non-synchronous time of sampling or local hydrologic cycle variations. But three localities, Winslow Water Company, Heislerville, and Long Beach Island, show the least correlation with the others. While the first is perplexing, the others might be the result of their coastal location.

There seems little correlation between each monitoring well and the annual dryness based on the available moisture index calculated from the weather data recorded at Indian Mills. This too may be the result of local hydrologic conditions or problems with seasonal high water table sampling time.

Finally, even with the period of record extending back to 1936 (i.e., Penn State Forest), this is probably still too short a duration to document longer-term geological phenomena.

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Periglacial Sediment-Filled Wedges, Northern Delaware, USA

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ABSTRACT

Wedge-like sedimentary structures have been observed at two sites in northern Delaware USA. The wedges, located along the erosional surface of and extending into channel deposits of the fluvial, mid-Pleistocene Columbia Formation, are 0.25-0.60 m wide at the top, 1.0-1.5 m in vertical extent, contain moderate to poorly sorted, vertically stratified sand, and are overlain unconformably by a layer of wind-blown silt. Several mechanisms, each with its own set of requisite environmental conditions, could hypothetically be responsible for the formation and infill of the wedges. Detailed physical, stratigraphic, and sedimentological information was used to evaluate competing hypotheses of wedge formation and sediment transport and eliminate those mechanisms that could not have formed the wedges. The best explanation for the Delaware wedges is that they are relict cryogenic structures formed by thermal contraction cracking in permafrost during the coldest parts of the Wisconsinan glaciation.

INTRODUCTION

Relict periglacial features associated with Late Quaternary glaciations have been used to reconstruct the periglacial environment south of the glacial border (Brown and Péwé, 1973; Washburn, 1980; Péwé, 1983). Relict periglacial features and the environment in which they formed are well known in the Appalachian Highlands of the eastern United States (Clark and Ciolkocz, 1988), but little is known about the extent, timing, and intensity of periglacial conditions affecting the coastal plain of New Jersey, Delaware, and Maryland. Features believed to have formed under permafrost or periglacial conditions in Mid-Atlantic region include sedimentary structures resembling ice-wedge casts (Walters, 1978; Newell et al., 1988), cryoturbated soils, and shallow depressions interpreted as thaw-lake basins, pingo scars, and deflation hollows (French and Demitroff, 2001). In this paper we discuss wedge-like sedimentary structures, referred to here as *sediment-filled wedges*, located in northern Delaware.

LOCATION AND GEOLOGIC BACKGROUND OF WEDGES

Sediment-filled wedges were observed within the Columbia Formation, a massive fluvio-glacial deposit, at two locations in New Castle County, Delaware (Figure 1). The first, a state-owned sand and gravel pit located south of Middletown at 39.44° N latitude and 75.71° W longitude, contains numerous wedges of varying sizes along pit walls. The second site is a privately owned gravel pit located northeast of Kenton at 39.25° N latitude and 75.64° W longitude that contains only one wedge.

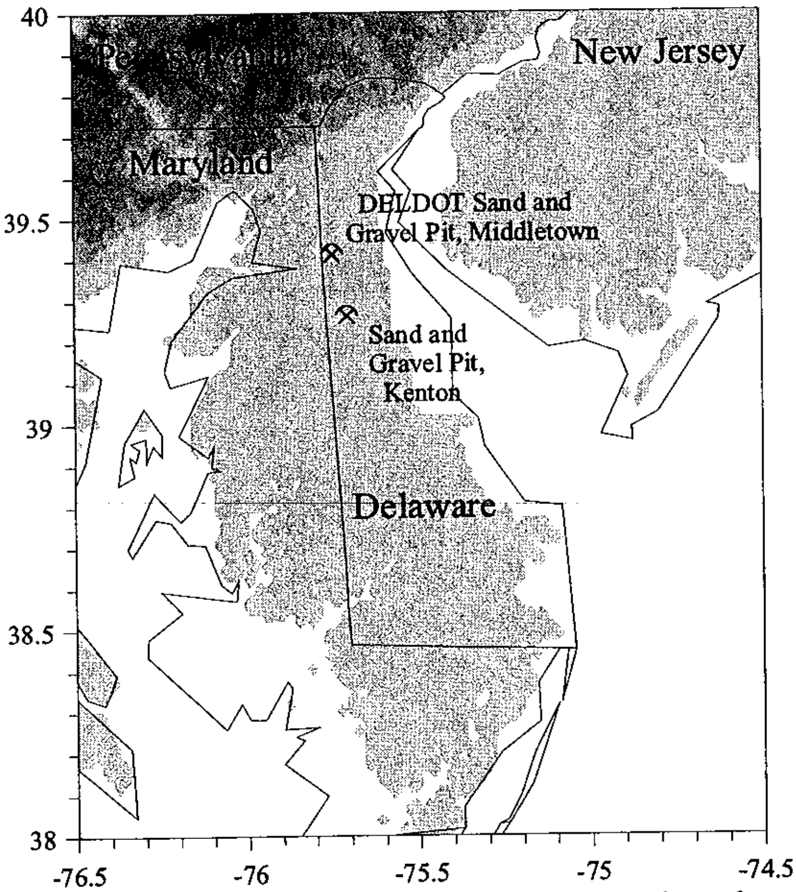

Figure 1. Map of northern Delaware showing the locations of Middletown and Kenton sand and gravel pits

Sediment-filled wedges near Middletown (Figure 2) and Kenton (Figure 3) are 0.2 to 0.6 m wide and extend subvertically 1 to 1.5 m, although larger wedges have been observed at the Middletown location. The top of each wedge occurs along the boundary between the Columbia Formation, deposited prior to 430,000 years before present (Groot and Jordan, 1999), and an overlying layer of wind-blown silt, approximately 13,000 years old (Ramsey and Baxter, 1996), at an average depth of about 1.0 m at Middletown and 0.6 m at Kenton. Wedges are spaced approximately 10 to 30 m apart but their three-dimensional geometry is unknown.

Surrounding layers of sand and fine gravel in the Middletown quarry are either undeformed or downturned slightly along the edges of each wedge. The cross-bedded layers of sand within the Kenton sand and gravel pit are slightly (1 cm) upturned along the edges of the wedge observed at that site. Within the Middletown quarry, layers of gravel have been offset as much as 8 cm along the edges of several wedges. Sediments filling the wedges at both sites consist primarily of vertically stratified, medium sands that are mineralogically similar to, but significantly more rounded, than that of the surrounding Columbia Formation.

Figure 2. Photograph of a sediment-filled wedge located within the DELDOT sand and gravel pit south of Middletown, Delaware.

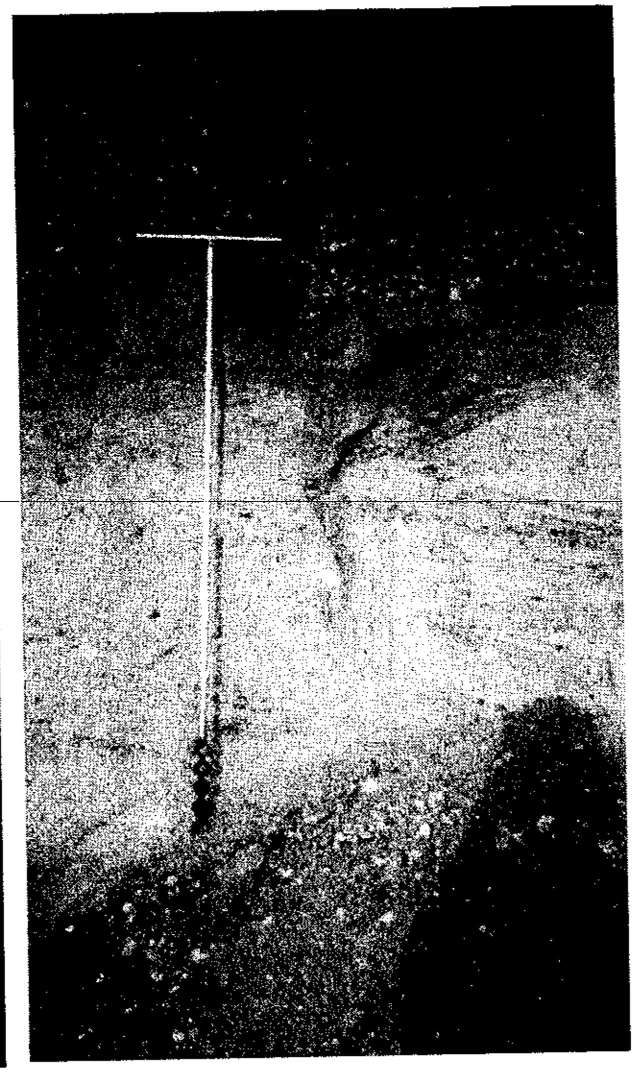

Figure 3. Photograph of a sediment-filled wedge located within a privately owned sand and gravel pit near Kenton, Delaware.

FORMATION OF SEDIMENT-FILLED WEDGES

Sediment-filled wedges can form by thermal contraction cracking (Black, 1976), in response to non-thermal tension within the ground (Black, 1976; Murton, et al., 2000), by desiccation cracking (Paik and Lee, 1998), or through seismic activity (Murton, 1996). Cracks are subsequently filled with sediment, creating wedge-like structures with sedimentological properties dissimilar to those of the surrounding material. This study uses Chamberlain's (1897) Theory of Multiple Working Hypotheses to determine the mechanism responsible for the initial cracking and subsequent infill of the sediment-filled wedges in northern Delaware.

Non-Periglacial Origins

Sediment-filled wedges may form in response to changes in temperature and precipitation, as well as near-surface geological processes. Detailed physical, stratagraphic, and sedimentalogical data collected from the Delaware wedges were compared to descriptions of wedges formed in response to nonthermal processes including nonthermal tension cracks, kettle cracks (Black, 1976), desiccation cracks (Paik and Lee, 1998), small faults, water escape structures (Murton et al., 2000), and soil fingers (Hinkel, 1993).

After evaluation of such competing hypotheses, it was determined that none of the non-periglacial processes included in this study could have been responsible for the initial cracking and subsequent infilling of the Delaware wedges. It was also determined that they were not altered after formation by the process of wetting and drying.

Periglacial Origins

Sediment-filled wedges may also form as the result of thermal contraction cracking within perennially frozen ground and can be filled with water, sediment or both. Wedges formed in this manner are referred to respectively as ice, sand, and composite wedges (Black, 1976; French, 1996). Sediment-filled wedges formed by thermal contraction cracking in seasonally frozen ground, including the active layer, produce frost wedges, which, like their permafrost counterparts, can be filled with water or sediment (Svensson, 1977; Dijkmans, 1989).

The Delaware wedges have characteristics similar to sand wedges formed within permafrost but in some ways resemble frost wedges formed in seasonally frozen ground. Dylik and Maarleveld (1967) and Murton et al. (2000) each list a set of seven criteria for distinguishing wedges formed under permafrost conditions from those of non-permafrost oringins. Of the seven criteria for wedges formed within permafrost outlined by Dylik and Maarleveld (1967), five, possibly six (large-scale polygonal pattern), are present in the Delaware wedges. Of the seven criteria compiled by Murton et al. (2000), four, possibly five (large-scale polygonal network), are present in the Delaware wedges. Thorough comparison of the Delaware wedges to those formed under permafrost and non-permafrost conditions indicate that the sediment-filled wedges in northern Delaware are most likely sand wedges formed within permafrost.

EVIDENCE FOR COLD CLIMATE ORIGIN OF DELAWARE WEDGES

The former presence of permafrost as far south as Delaware cannot be confirmed without other evidence such as identification of nearby relict features of similar age and origin. Relict ice wedges and other periglacial features located in central Pennsylvania (Newell et al., 1988) and northern New Jersey (Walters, 1978) indicate that permafrost did exist within the Mid-Atlantic region. Several geomorphic features, including dunes, sand sheets, shallow depressions, and folded, contorted, and disrupted bedding, have been observed at several locations in Delaware and, based on stratigraphic, sedimemtological, and palynological evidence, are believed to have formed during the coldest parts of the Wisconsin (Andres and Howard; 1998; Ramsey, 1998). Wedge-like structures, similar to those described in this paper, have been observed within the Scotts Corners Formation near Dover, Delaware (Andres and Howard, 1998), the Bridgeton Formation in southern New

Jersey (Newell et al., 1988), and have been described elsewhere along the New Jersey coastal plain by French and Demitroff (2001).

Pollen assemblages for northern Delaware indicate that a taiga or tundra-like environment did exist in Delaware as the Laurentide Ice Sheet reached its southernmost exent (Sirkin et al., 1977; Colman et al., 1990). Pollen data, in conjunction with other probable cold-climate features, indicate that a periglacial environment did exist in northern Delaware during last glacial period, providing a suitable environment for the formation of the primary sand wedges near Middletown and Kenton.

CONCLUSION

The sediment-filled wedges in northern Delaware are morphologically and sedimentologically similar to wedge-shaped structures formed by thermal contraction cracking within permafrost. Although it cannot be stated conclusively that permafrost did exist in Delaware the morphology and size of the sediment-filled wedges along with other paleoenvironmental information are indicative of its presence. The exact time in which the wedges formed has yet to be determined, but based on ancillary evidence we can infer that they formed at the peak of the last ice age. After the initial crack opened it was subsequently filled with aeolian sand, ceasing formation prior to deposition of the overlying layer of silt around 13,000 years before present. A more detailed description of this study is provided in Lemcke and Nelson (in review).

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Paleoclimatic Implications of Blockfield Distribution in the Central Appalachians

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ABSTRACT

The NE-SW trending Appalachian Mountains extend nearly 2000 km in the eastern United States, and contain many relict periglacial features such as blockfields, ice wedges, and sorted patterned ground. Eleven groups of blockfields were identified, using published evidence in conjunction with air photos and USGS topographic maps. The groups are arrayed along a transect in the Appalachians, with endpoints in central Pennsylvania and North Carolina. Information about the location, elevation, slope, aspect and size were collected for each blockfield. A DEM-aided interpolation routine was used to estimate modern air temperatures for each blockfield location. Exploratory data analysis has revealed a general decrease in the median elevation of blockfields with increasing latitude. This trend is consistent with a hypothesized climatic (periglacial) origin for many Appalachian blockfields.

INTRODUCTION

The NE-SW trending Appalachian Mountains extend nearly 2000 km in the eastern United States, and contain many relict periglacial features (Péwé, 1983; Clark and Ciolkosz, 1988; Clark and Schmidlin, 1992). The terms *blockfield* and *boulder field* have been used to describe a specific type of periglacial landform comprised of large areas of cobble to automobile-size clasts the; term "blockfield" will be used here to describe such features. Research on relict Appalachian blockfields indicates that their formation was dependent to some extent on local geology and climatic conditions. However, the formation of periglacial blockfields requires an intensely cold climate to prevail over an appreciable period. Such conditions existed in the Appalachian Mountains during the Pleistocene, resulting in the formation of numerous blockfields that are still evident today.

BACKGROUND

Block deposits are among the most peculiar and spectacular landforms in the Appalachians. Block deposits typically involve large unvegetated areas containing cobble to boulder-sized clasts. It has been established through multiple lines of evidence that block deposits (excluding scree slopes) in the Appalachians are inactive today. Evidence supporting this view includes undisturbed forest growth on and adjacent to block deposits, secondary differential weathering of blocks, lichen growth on exposed block surfaces, and accumulation of organic-rich soil on and between some blocks (Middlekauff, 1991). The resistance of quartzitic sandstone to weathering and the blocky armor of block fields make it plausible that such features are periglacial relicts that exceed 10,000 years in age.

Although block deposits are widely attributed to intense frost action and deposition, the particulars of the transport process are still vague due to a lack of modern analogs. There is substantial evidence to suggest that many Appalachian boulder fields are periglacial deposits created by intense freeze-thaw activity accompanied by periglacial mass-movement processes (e.g., Smith 1953; Potter and Moss, 1968). Evidence to support this hypothesis includes: (1) well jointed escarpments at the upslope margin of boulder fields; (2) the progressive decrease in boulder size and angularity with increasing distance from scarps; and (3) segregation and vertical orientation of larger boulders on the surface of block deposits, indicating that thaw settlement accompanied degradation of ice-rich permafrost. The high density of large, well-developed block fields in close proximity to the Wisconsin glacial border also supports a periglacial interpretation.

METHODOLOGY

The Appalachian Mountain Range is a narrow, linear physiographic feature that lends itself well to studies of climatic elements at various elevations along a latitudinal transect. Using the *TerraServer* imaging tool (www.terraserver.com, 2000), 11 groups consisting of 96 individual blockfields were identified along a transect with endpoints close to the glacial border in Pennsylvania and the Smoky Mountains near the North Carolina-Tennessee border. Only unvegetated blockfields were considered in this study, as these could be identified easily on aerial photographs. Field verification of several sites was conducted using GPS and USGS topographic maps. The distribution of blockfield groups is shown in Figure 1.

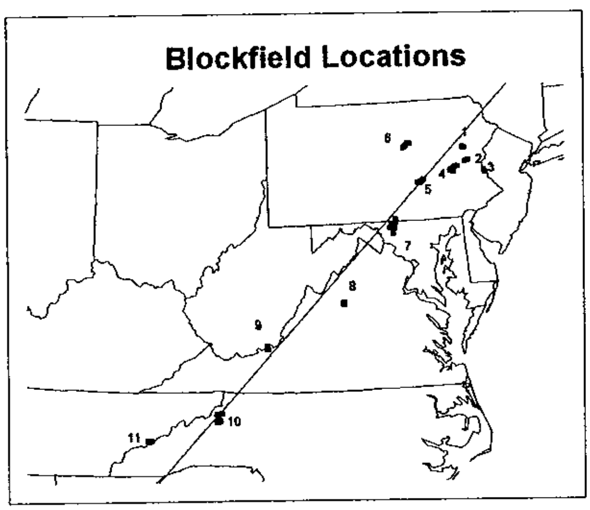

Figure 1. Geographic distribution of blockfields (11 groups). Solid line shows the approximate course of transect over which blockfields were sampled.

The median elevation of blockfields increases with decreasing latitude. This is significant because mean temperatures generally increase with decreasing latitude in the Northern Hemisphere. To examine this relation further, the latitude and elevation of each blockfield were plotted. Figure 2 depicts the median, maximum, and minimum elevation of each blockfield group. A clear decrease in the minimum elevation of blockfield groups occurs with increasing latitude. This trend is hypothesized to be the result of a climatic gradient along the spine of the Appalachians.

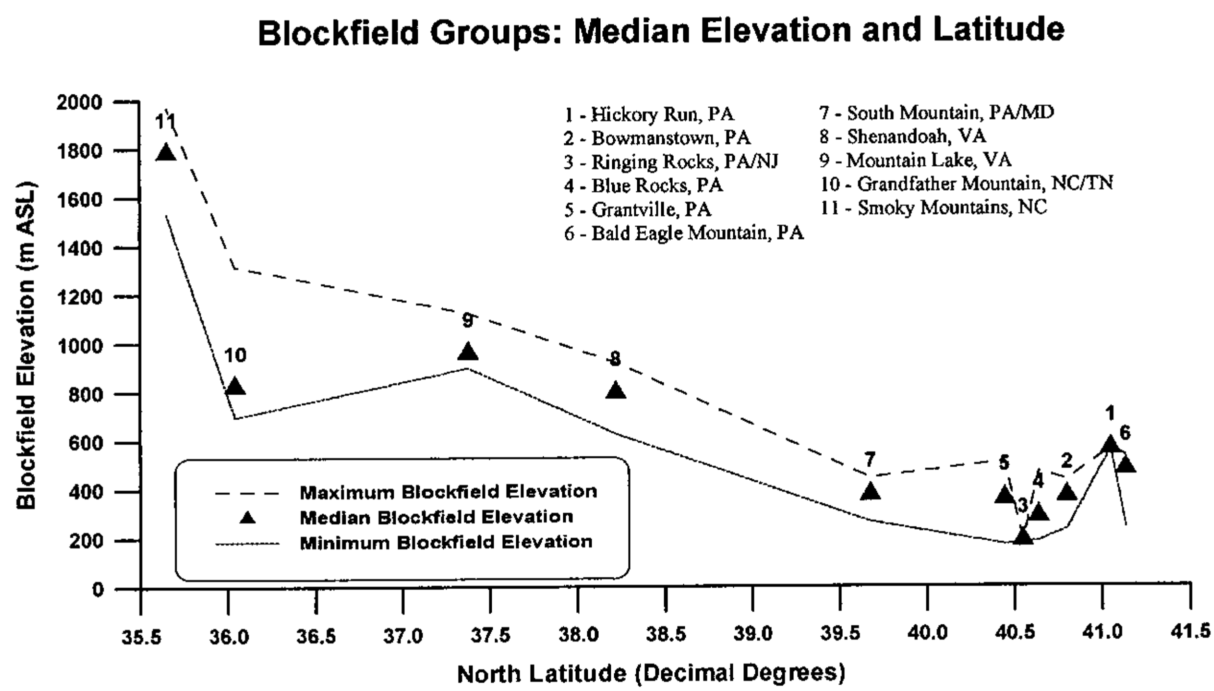

Figure 2. Elevation trends of blockfields in the Appalachian Mountains. including median, minimum, and maximum elevation of each group.

ANALYSIS: ESTIMATING AIR TEMPERATURE

Ten-year means of National Climate Data Center summer air temperature data (NCDC, 1999) from 64 cooperative weather stations located throughout the Appalachians were used in a DEM-aided inverse-distance-squared interpolation routine (Willmott and Matsuura, 1995; Tveito and Forland, 1999) to estimate modern temperatures at each blockfield derived through lapse-rate calculations (Leffler, 1981; Walegur and Nelson, 2000). The mean summer temperature for each blockfield is plotted vs. latitude in Figure 3. The blockfields all occupy points that experience similar modern climates with respect to temperature. Ten of the eleven blockfield groups range from 18-22°C. The consistency of this estimate between sites suggests that even widely separated blockfields experienced similar climates during the Last Glacial Maximum when temperatures in the Central Appalachians were lowered sufficiently to drive periglacial processes. All seven of the blockfields outside this range are in Group 11, in the Smoky Mountains of North Carolina/Tennessee. This is the southernmost group and contains the highest elevation blockfields in the study area, with a median elevation of 1784 m.

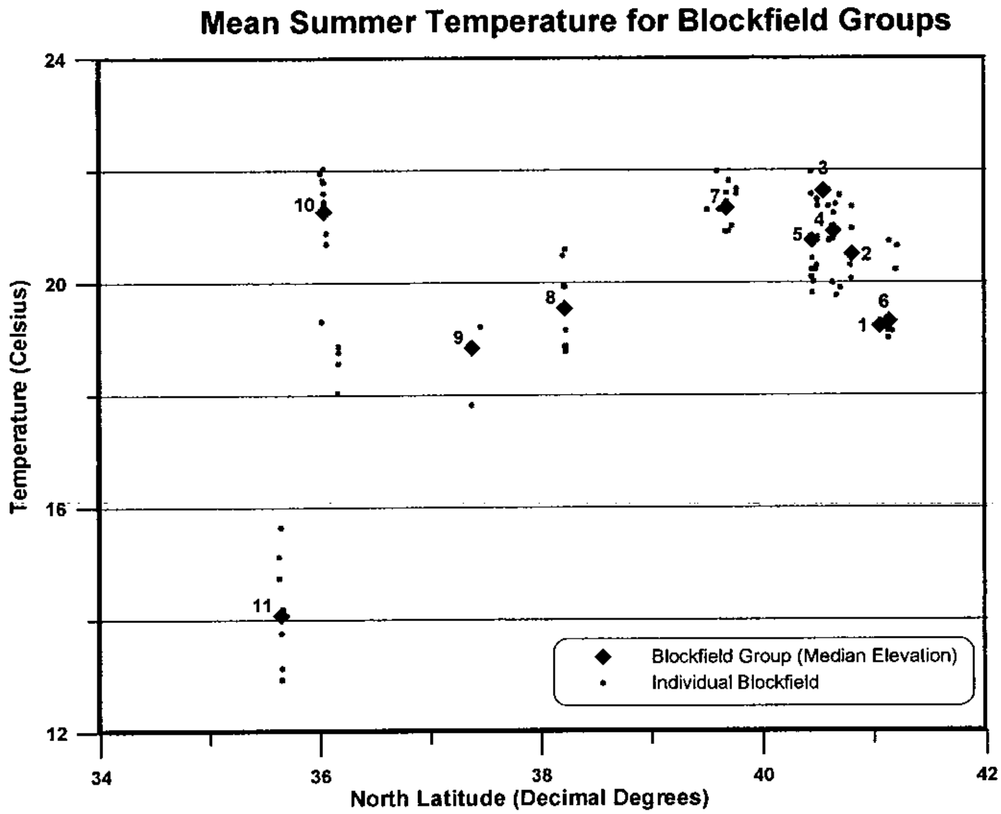

Figure 3. Estimates of modern mean summer air temperature at blockfield locations.

CONCLUSION

The distribution of relict periglacial blockfields in the Appalachians appears to be related to paleoclimatic temperature trends. Current mean summer temperatures in blockfield areas were calculated to be 13-22°C. This range of temperatures is smaller (approximately 18-22°C) for 89 of the 96 individual blockfields. The altitudinal trend and narrow range of current temperature estimates over the Appalachian transect indicate that the blockfields have a periglacial origin. Blockfields at lower latitudes generally exist at higher elevations, but in similar temperature regimes.

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Spung Map: Great Egg Harbor River Watershed Region, Southern New Jersey

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ABSTRACT

Evidence is mounting that two intervals of semi-polar desert conditions once existed in southern New Jersey during the Late-Pleistocene (French and Demitroff, 2001; French et al., 2003; Demitroff, this volume; French and Demitroff, this volume; also see Lemcke and Nelson, this volume). Periglacial features, such as cryoturbation and ancient sand wedges have been discovered in southern New Jersey and into southern Delaware, suggesting the existence of discontinuous permafrost and deep seasonal frost. Ventifacts found throughout southern New Jersey indicate that wide-spread eolian processes were operating. Strong winds blowing across sparsely vegetated, sandy-gravelly "weak zones" in the frozen landscape allowed for the formation of deflation hollows. Following glacial retreat at the end of the Pleistocene, the ground thawed and temperate conditions returned to the region in Early Holocene time. In response to glacioeustatic sea level fluctuations and abundant glacial melt water, aquifers recharged and ground water tables rose to intersect the hollows, forming enclosed wetland basins and vernal ponds. In southern New Jersey, these features are colloquially termed "spungs".

Spungs carry significant geologic, biologic, archeologic and environmental importance, and may serve as sensitive indicators of hydrogeologic change. Many have dried and disappeared during the past century, while others remain waterless for greater periods of time (see Demitroff, this volume). To facilitate analysis of change and monitor the status of these features, GIS-RS-GPS (geographical information system, remote-sensing, and global positioning system) methods were used to catalog and create a digital map of spungs in the Great Egg Harbor River watershed (Fig. 1). Spung outlines were delineated and georeferenced using 1930-32 aerial photographic imagery (obtained from Mark Demitroff through the Great Egg Harbor Watershed Association, personal communication, 2003) and 1995-97 aerial infrared imagery (New Jersey Department of Environmental Protection, 1998). The 1930-32 imagery represents a 4th generation product, scanned as tiff images at 1,200 dpi from 8-by-11 inch internegatives (3rd generation), photographed from 1:12,000-scale 3-by-4 foot prints (2nd generation), which in turn were derived from manually-spliced 9-by-9 aerial photographic negatives (1st generation) (Mike Ryan, Aerial Photo and Historic Map Library, Bureau of Tideland Management, New Jersey Department of Environmental Protection, Personal Communication, 2001). The 1930-32 images were georeferenced to the 1995-97 imagery. The spungs were delineated on the georeferenced 1930-32 images, and their positions further refined using the 1995/97 imagery. A sampling of delineated spung outlines were ground-truthed with GPS to confirm their extent and positions.

The spung map identifies 40 features of various size, shape, and configuration that are known to have names. The spungs occur as separate, lone entities, or in clusters. Separate spungs are generally larger in size than the individual spungs that make up a spung group.

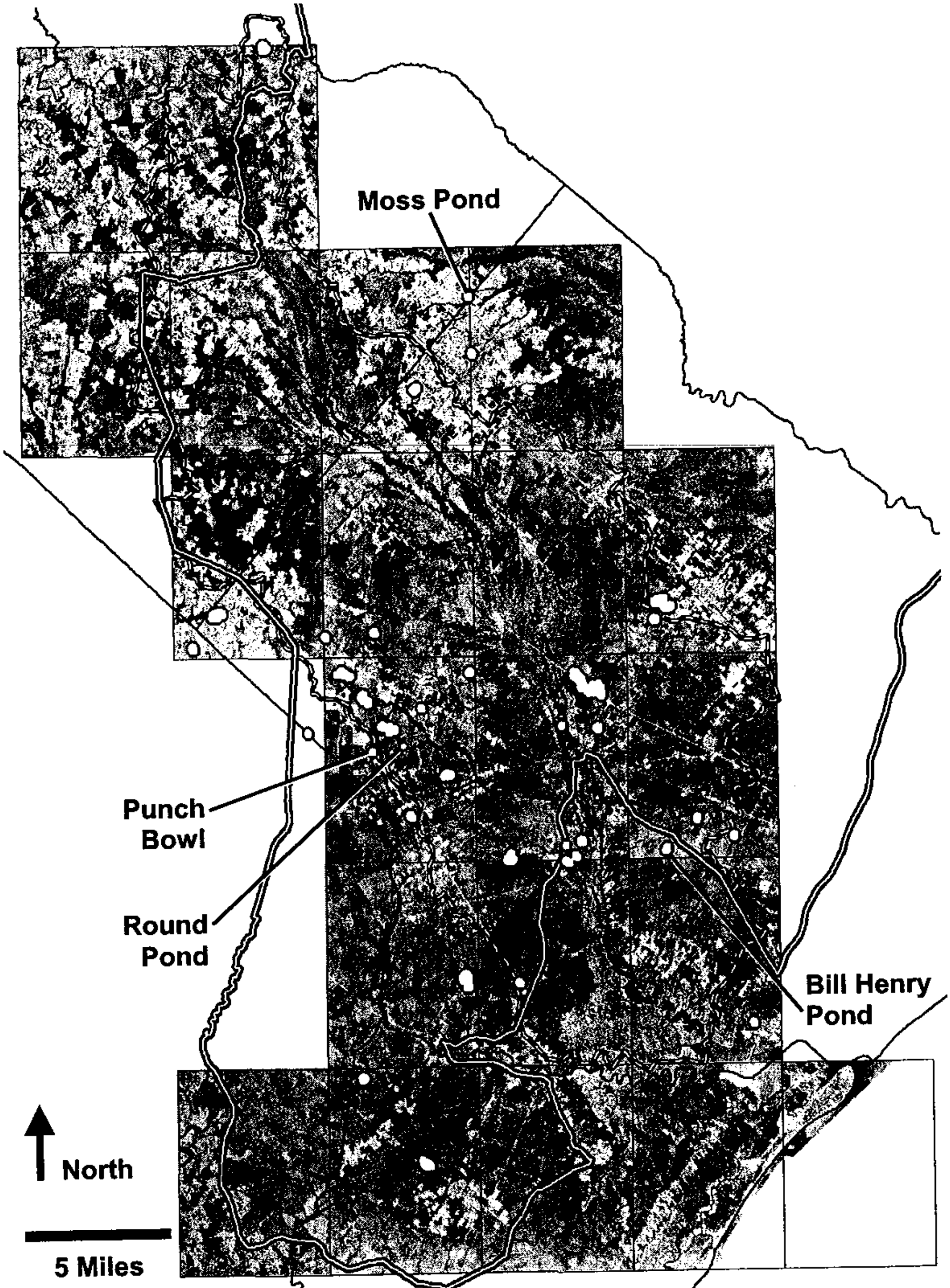

Figure 1. Spung map of the Great Egg Harbor River watershed. Mapped spungs are shown as white polygons with black outlines (grossly exaggerated in size so as to be visible) overlain on 1930-32 aerial photographic images (image tiles shown as black-outlined boxes). The spungs extend throughout the watershed (dashed black-white line), the Pine Barrens (black-outlined white line), and parts of Atlantic, Camden, Cape May, Cumberland and Gloucester counties (thin black line). Named spungs are discussed in the text.

The separate spungs, which number thirty-three, range in size (plan-view surface area) from 72.8 to 0.14 acres, with a mean size of 12.9 acres (standard deviation of 16.0). Twenty-four individual spungs comprise seven spung clusters, and range in size from 15.1 to 0.06 acres, with a mean size of 2.3 acres (standard deviation of 3.6). When the individual spungs within a cluster are considered as a whole, the forty spungs have a mean size of 13.4 acres (standard deviation of 15.2).

A visual comparison between the 1930-32 and 1995-97 imagery indicates that some spungs have persisted as identifiable features, whereas others have completely disappeared. Three scenarios were noted: (1) no significant difference in spung outline, (2) some degree of change visible, and (3) no evidence in the 1995-97 imagery to indicate that a spung had existed. "Bill Henry Pond" is an example of the first condition, and appears very similar in size and extent in the 1995-97 image as it does in the 1930-32. "Moss Pond" is an example of the second scenario, and is seen as a water-filled pond amidst farmland in the 1930-32 image, but is visible only as wetlands in the 1995-97. The outline and position of the spung has remained the same, but there is no water visible. "Punch Bowl" is an example of the most extreme condition (see Demitroff, this volume, figure 12). Once a small area of wetlands visible in the 1930-32 image, it is now thickly overgrown and completely dry, and half of the spung has been in-filled and plowed-under as part of a farm field. The spungs that have clearly persisted as visible features are located in isolated areas or regions that have experienced minimal development. Conversely, spungs that are not as readily identifiable in the 1995-97 imagery, particularly those that have been overgrown by wetlands or forest, are in areas that have seen significant increases in development and agricultural activity since the 1930s. Furthermore, the size of a spung (as delineated in the 1930-32 imagery) appears to have little impact on its persistence. "Round Pond", a separate spung and the smallest delineated, remains visible and well defined on the 1995-97 image. It is located in an area of little development. These findings are consistent with the observations made by French and Demitroff (2001) that increased urbanization and agriculture withdrawal may have resulted in local and(or) regional lowering of ground water table, as manifest by the change in appearance and visibility of some spungs in the 1995-97 imagery.

Based on the initial observations above, the spung map and subsequent additions will be useful tool to help assess ground water in southern New Jersey. Further spatial pattern analysis of spung shape, orientation, and distribution, as well as GIS-based analysis of stream valley asymmetry in southern New Jersey, will shed light on the eolian and periglacial processes that once operated in this region.

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Vernal Pools in New Jersey's Outer Coastal Plain

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ABSTRACT

Vernal pools are confined wetland depressions, either natural or man-made, that typically hold water between fall and mid-summer and are devoid of breeding fish populations. Vernal pools are widely distributed throughout rural regions of New Jersey, with the largest concentration [5,834 of 13,559 (43%)] occurring in the Outer Coastal Plain geophysical province. Outer Coastal Plain vernal pools vary in size from $5m^2$ to >1ha. These ephemeral wetlands support rich amphibian and reptile species assemblages, which are generally higher in diversity than many permanent bodies of water. Amphibians that breed in vernal pools are classified into two groups defined by their reliance on these wetlands for the completion of their life cycle: *obligate* – species that breed exclusively in vernal pools, and *facultative* – species that use both permanent and temporary wetlands. New Jersey possesses five species of *obligate* vernal pool amphibians, three of which occur in Outer Coastal Plain pools: *Ambystoma opacum* (marbled salamander), *Rana sylvatica* (wood frog) and the state-endangered *Ambystoma t. tigrinum*, (Eastern tiger salamander). *Rana sylvatica* is the one of the most abundant amphibians in Outer Coastal Plain vernal pools. In 2001 the New Jersey Department of Environmental Protection adopted legislation to afford protection vernal pools that met specific physical and biological criteria, including a two-month minimum hydroperiod, hydrological isolation from other wetlands, and the documentation of breeding by at least one obligate or two facultative amphibians. Vernal pools that meet these criteria are classified as ***certified***. The vernal pool protection legislation coincided with the Division of Fish and Wildlife's volunteer-based Vernal Pool Survey Project, which to date has certified 675 vernal pools statewide, 180 which are located in the Outer Coastal Plain.

New Jersey – Under the Ice

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ABSTRACT

This PowerPoint presentation is available at: *http://hurri.kean.edu.* Click on the icon – *Geology of New Jersey.* (One slide has sound).

Knowledge of Pleistocene glaciation in approximately the northern one-quarter of New Jersey comes from the effects the glaciers had on the landscape in the deposits they left behind and to lesser degree in the re-shaping of the pre-glacial topography. The major deposits consist of the terminal and recessional moraines, several varieties of till, various outwash deposits, drumlins, eskers, and lake bottom sediments. Smoothed and reshaped hills and valleys, erratics, waterfalls and numerous lakes, ponds, and swamps (which otherwise would not exist) are also evident.

The most prominent moraine is the Wisconsin Terminal Moraine that extends from Perth Amboy in a sinuous stretch across the state west to Belvidere. In some places it is well expressed topographically but in others is scarcely noticeable. It goes by various local names, as the Perth Amboy, Madison, Budd Lake, and others. The terminal moraine marks the demarcation line between glaciated and non-glaciated terrain. The glaciated area north of the moraine is characterized by an abundance of lakes, ponds and swamps and waterfalls that are absent south of the moraine. Other terminal moraines, less well preserved and south of the Wisconsin Terminal Moraine, indicate earlier glaciations.

Till of various types occurs extensively throughout the glaciated area though it may be overlain by other deposits. Outwash deposits are common either as pro-glacial outwash plains or a variety of lake-deltaic and glacio-fluvial deposits.

Drumlins are common and as streamlined forms indicate directions of ice movement. Several eskers are present and present typical sinuous topographic forms.

Glacial lakes were common, 33 sufficiently in evidence to warrant names, and a variety of deposits and landforms resulted - deltas, shorelines, wave-cut terraces, and outlet channels. The most prominent lake was Lake Passaic that extended north-south from Far Hills to Pompton Plains and was about 8 miles wide east-west. Its extent varied as the ice first advanced and blocked outlets and raised the lake level, and then retreated and opened outlets that lowered the lake level. A prominent outlet formed at Moggy Hollow in the town of Far Hills and is a prominent topographic feature.

The ice stood about 2000 feet high above High Point in northernmost New Jersey, the highest hilltop in New Jersey at 1805 feet. The direction of ice movement is indicted by striations on bedrock surfaces and they indicate changing directions in many areas.

Software Developed for Near-Real-Time Internet Seismic Signals

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Moravian College
and
Michael J. Sands
Essent Corporation

ABSTRACT

A software system, named Seismic Internet Monitoring Application (SIMA), has been developed to allow the near-real-time display and evaluation of seismic traces on the Internet. SIMA consists of three components: (1) an embedded microsystem (called a TINI) that is connected to as many as four seismometer amplifiers, (2) a network-enabled server that displays and broadcasts the seismic signals over the Internet, and (3) a client software application that allows for remote display, evaluation, storage/retrieval of the seismic data. Presently, seismic signals are available from two institutions, Moravian College and Kean University, to demonstrate the functionality of the system. All SIMA software components are free and are available for downloading from http://www.physics.moravian.edu/seismic. Any person with a computer that runs any version of Windows from 95 through XP can receive near-real-time seismic signals from these sites over the Internet. Institutions with seismometers need only the TINI hardware (cost: approximately $250) to broadcast their signals over the Internet via SIMA software.

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Late Pleistocene Periglacial Phenomena in the Pine Barrens of Southern New Jersey: GANJ Field Excursion Guide, October 11, 2003

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University of Ottawa
and
Mark N. Demitroff
Buckhorn Garden Service Inc.

INTRODUCTION

The excursion aims to demonstrate Late-Pleistocene periglacial phenomena in southern New Jersey. Evidence is also presented that indicates permafrost conditions prevailed in southern New Jersey on at least two different occasions during the Late Pleistocene.

Periglacial environments are those in which frost-action and permafrost-related processes dominate (French, 1996, p3). Permafrost is perennially frozen ground. The near-surface layer of permafrost terrain is subject to seasonal thawing and freezing; this is termed the active layer.

The Late-Pleistocene was a period of the Quaternary when world climates and global sea levels fluctuated. In the northern hemisphere, large continental-scale ice sheets developed, and episodes of cold, often arid, conditions characterized many of the unglaciated mid-latitude regions. A so-called 'periglacial zone' formed in areas peripheral to the ice sheets. Southern New Jersey was part of the Pleistocene periglacial zone of eastern North America.

The excursion emphasizes the Pine Barrens, an area of generally flat, sandy and gravely terrain that extends through southern New Jersey. The higher elevations rarely exceed 45 m above sea level. The regional groundwater table is usually less than 1.5 m beneath the ground surface. Much of the area is forest re-growth that dates from the 1930's. The area was likely given its name by early settlers who found the sandy soils generally unsuitable for cultivation.

THE PINE BARRENS

The New Jersey Pinelands National Reserve (figure 1) comprises approximately 1.4 million acres, approximately 20% of the state of New Jersey. It represents the most extensive wilderness tract along the Mid-Atlantic seaboard. Ecologically, the area consists of generally flat terrain with sandy and acidic soils. The underlying sediments consist of thick sequences of silt, sand, and gravel of Cretaceous and Tertiary age. The surface is mantled either by light-colored 'sugar' sand, or by wind-blown silty sand that forms shallow dunes, 1-3 m high. Depressions form wetlands and many contain shallow or ephemeral ponds. A few deeper water bodies probably reflect the sites of groundwater upwellings, or subterranean springs. Many pond levels fluctuate seasonally, indicating groundwater recharge, and some of the wetlands are ephemeral.

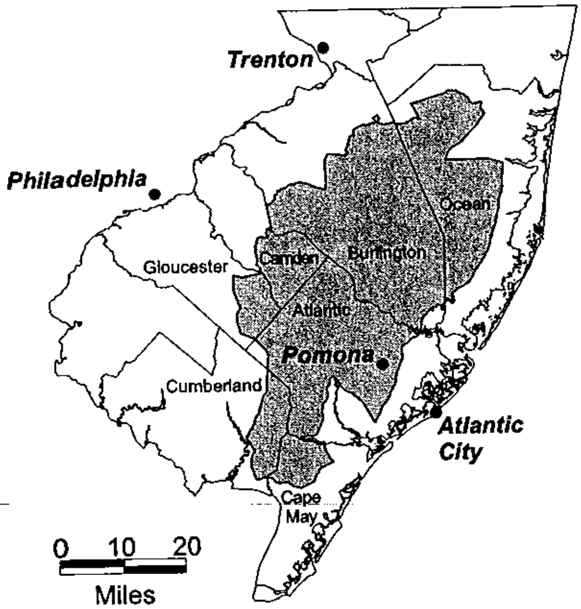

Figure 1. Map showing extent of Pinelands National Reserve.

The forest consists of unbroken pitch pine and various oaks. Atlantic white cedar and deciduous broad leaf trees grow in the more humid soils.

Drainage of the Pine Barrens is via the Great Egg Harbor, Mullica, Maurice and Toms rivers together with a number of relatively small, low-gradient streams that follow large shallow valleys. The streams are characteristically acidic and nutrient poor. The action of chemotrophic bacteria oxidizes soluble iron in the groundwater. This flocculates out as bog iron (Crerar et al, 1979). Natural organic substances combine with the Fe to produce a dark tea color to the waters of the area.

The depressions and basins of southern New Jersey (figure 2) are known locally as 'spungs' (old English: pocket, purse), a term that probably relates to their profitable early use as cranberry picking localities. Others believe they should be referred to as 'spongs' (old English: long strip of swampy land; see Oxford English Dictionary, 1971; McPhee and Curtsinger, 1974). Closely related is the term 'cripple' (Oxford English Dictionary, 1971; Weygandt, 1940), a local word that refers to a dense thicket in swampy or low-lying ground. Today, the depressions form environmentally-sensitive wetlands in an area of predominantly sandy soils and limited surface water. A concern is that many of these wetlands are drying up, probably in direct response to a general lowering of the groundwater table as urbanization, commercial agriculture and groundwater usage increases.

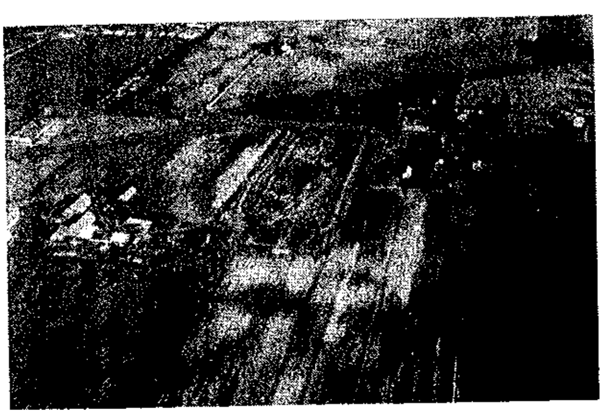

Figure 2. Examples of wetlands ('spungs') in the Pine Barrens of southern New Jersey.

A. Egg Harbor Pond, Vineland, Cumberland County, has been dry for many years and is now under cultivation. The outline of the depression can be seen in the centre of the photograph. Photographs from 1930-32 indicate that a pond existed at that time. Photo was taken in March 2001.

B. Horsebreak Pond, near Buena, Atlantic County, has undergone substantial drainage during the last 25 years. At the time this photograph was taken (March 2001), the swamp was dry.

EXCURSION ROUTE

The excursion begins at Stockton and proceeds north via Hammonton towards Vincentown (Stop 1) and Mt Laurel (Stop 2). It then returns south through Lebanon State Forest (Stop 3). After lunch, the excursion makes a north-south transect through the Pine Barrens (60+ minutes by bus) to Port Elizabeth (Stop 4). The return to Stockton is via Newtonville (Stop 5) and Hammonton.

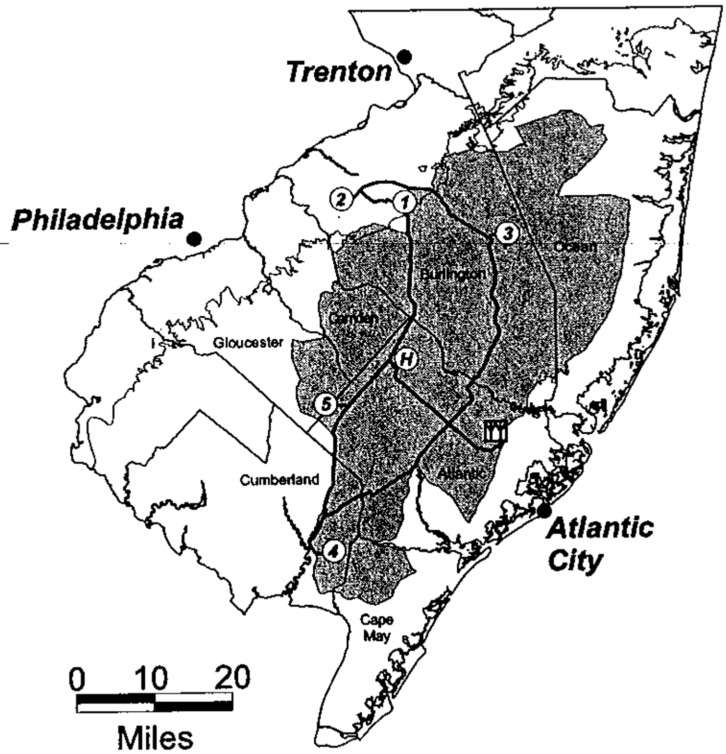

Figure 3. Excursion route, October 11, 2003. Stops 1-5 are indicated. The letter "H" denotes the Hammonton Square Mall. The Stockton logo shows the location of the College (small square with two side-by-side trees).

DETAILED ROAD LOG

The road log begins and ends at Hammonton Square, Hammonton.

Log (miles)	Direction
0.0	Depart Hammonton on Highway # 206 north.
20.7	Left on Main Street, Vincentown
21.4	Left on Mill Street
21.6	Right on Landing Street.
21.8	Walk 150' in field on right. STOP # 1
21.8	Continue on Landing Street
23.6	Left on Eayrestown Road
23.9	Right on Bridge Road

25.3 Left on Highway # 541

25.6 Right on Fostertown Road.

28.1 Left on Hainesport- Mt Laurel Road

29.6 PAWS Farm Nature Center. STOP #2.

29.6 Exit PAWS and turn northeast on Hainesport-Mt Laurel Road

31.6 Right on Highway # 36 east

40.5 Right on Magnolia Road (Highway # 644 east)

47.2 'Four Mile Circle'. Exit on Highway # 72 east

48.4 Left into Brendan Byrne State Forest

48.7 Right onto road to Pakim Pond. Pass Park HQ.

50.8 Left to Pakim Pond

51.3 Pakim Pond parking area. STOP #3. LUNCH

51.3 Exit Pakim Pond parking area. Turn right. Follow sign 'Exit park'

52.3 Turn east on Highway #72

52.7 Right on Highway # 563 towards Batsto. Pass Chatsworth, Jenkins Neck, Green Bank. Cross Mullica River. Continue south on Highway# 563.

79.6 Highway # 563 changes to Highway #50. Continue south.

87.2 Turn left at Mays Landing. Continue south on Highway # 50.

87.6 Right on Highway #669 west.

93.8 Right on Highway # 666 north.

94.4 Left on 13th Avenue going southwest towards Hesstown. Cross Tuckahoe, River.

100.3 Cross Highway #349. Continue south.

101.3 Left on Port Cumberland Road.

105.3 Left on Highway #47 south towards Port Elizabeth.

106.6 Turn onto Highway #347 south

107.1 Left into entrance of Mays Landing Sand and Gravel.

107.9 Gravel pit. STOP # 4.

108.9 Exit Mays Landing Sand and Gravel. Turn right on Highway #347 north.

109.2 Highway #347 ends and becomes Highway #47 north

110.5 Right on Port Elizabeth-Cumberland Road.

115.2 Left on Highway #49 west

115.7 Right on Union Road and proceed towards Buena.

125.0 Left on Tuckahoe Road (Highway # 557 north)

125.4 Stop sign. Keep in right lane and then left to Highway #49 west

125.5 Right on Highway #54 north

--- STOP # 5. Near Newtonville

--- Continue on Highway # 54 north towards Hammonton

133.3 Hammonton Square, Hammonton.

BACKGROUND GEOLOGY AND GEOMORPHOLOGY

The Quaternary Era

The last 3 million years of geological time is generally referred to as the Quaternary Era. It is subdivided into (1) the Pleistocene and (2) the Holocene (last 10,000 years).

The Quaternary Era was characterized by numerous and dramatic fluctuations of climate. During the colder periods, continental-scale ice sheets developed in the polar regions of the northern hemisphere. In North America, the maximum southern limits of the Late-Pleistocene ice reached as far south as northern New Jersey. The most recent glaciation, termed the Wisconsinan, reached its maximum extent approximately 18-24,000 years ago. There is evidence that two earlier Late-Pleistocene glaciations also reached their maximum southern extents in northern New Jersey. These are known as the Illinoian and Pre-Illinoian glaciations. The Illinoian glaciation is generally thought to have reached its maximum extent approximately 130-160 thousand years ago (oxygen-isotope-stage 6) while the Pre-Illinoian is older. Elsewhere, in both Europe and North America, there is evidence that even older glaciations occurred in the Middle and Early Pleistocene.

Periglacial conditions

During the glacial periods, the ice-free areas of the mid-latitudes experienced cold and generally arid conditions. The term 'periglacial' describes such environments.

Typically, the Pleistocene periglacial environments were characterized by (1) low mean annual ground and air temperatures, (2) the freezing and thawing of the ground, (3) the formation of either seasonally frozen ground or perennially frozen ground (i.e. permafrost), (4) the widespread occurrence of shrub-tundra, tundra or polar semi-desert conditions, and (5) strong winds. The latter were the result of the inevitable compression of the zonal westerly winds between the trade winds to the south and the high pressure systems that would have developed over the ice sheets to the north. In addition, local katabatic-winds would have dominated the immediate ice-marginal areas. The majority of Pleistocene periglacial environments were also arid and continental in nature. This was because global sea level would have been as much as 40-60 m lower than today. As regards southern New Jersey, the Atlantic Ocean would have been ~200 km further to the east. It is debatable whether modern terrestrial analogues exist for the Pleistocene periglacial environments (French, 2000).

The ending of the last periglacial episode in southern New Jersey was relatively short, probably lasting only 2-3,000 years. It was characterized by an enhanced level of geomorphic activity and landscape evolution. The amelioration of climate would have resulted in the thaw of permafrost and/or seasonally frozen ground. This process is termed thermokarst. The melt of ground ice would have resulted in especially rapid and dramatic landscape modification. Ground thawing would also have released sediment for transport, accentuated mass-wasting and slope instability, and allowed stream incision. At the same time, melting of the ice sheets would have allowed global sea level to rise, and the tundra and shrub-tundra vegetation of the periglacial zone would have been quickly replaced by trees.

Today's landscape

Southern New Jersey is a relatively flat, low-lying terrain with extensive areas covered by forests. Landforms are subtle and often difficult to recognize. The highest elements of the landscape are usually underlain by sandy gravel of the Bridgeton Formation, of Tertiary - early Pliocene age (Newall et al, 2000). The broad valleys and gentle slopes of the area are typically mantled with 1-3 m of Middle and Late Pleistocene sediment, mostly locally-derived. Fine wind-blown sands of Holocene age veneer the higher surfaces and alluvium accumulates in the lower elevations. Late-Holocene sea-level rise has resulted in the inundation of the lower stretches of the major river valleys and the formation of extensive coastal marshes and inter-tidal flats along the Great Egg Harbor, Mullica and other rivers, and along the immediate coastal environs.

A model of landscape evolution for southern New Jersey, presented by W. L. Newall et al (2000), is reproduced here as Figure 4. It indicates that an intimate relationship exists between surficial materials and landscape elements. The surficial materials are described as being the result of mass-wasting.

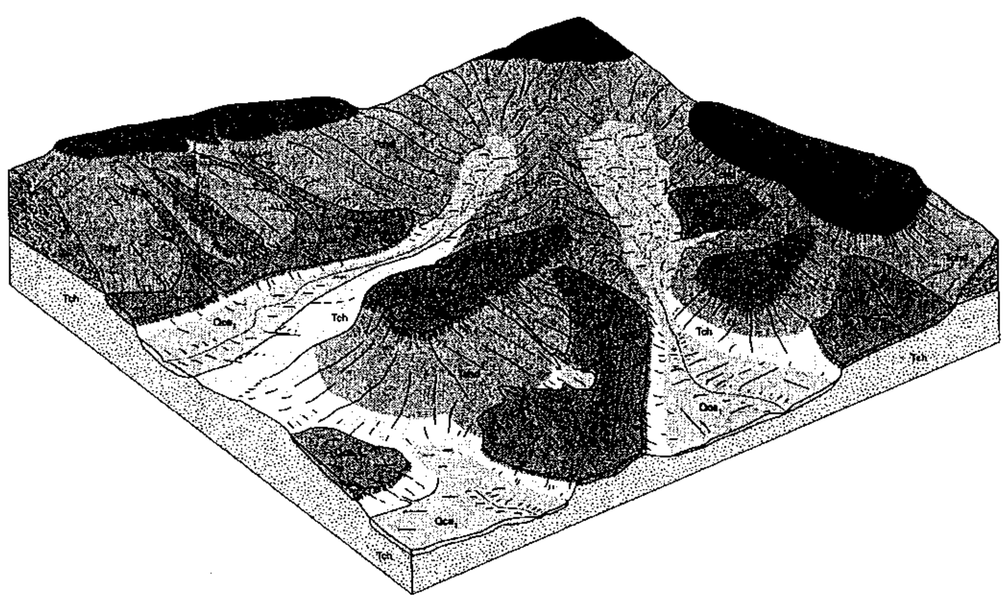

Figure 4. Block diagram showing schematic relationship between surficial materials and landforms in southern New Jersey. From Newall et al. (2000) figure 4.

Broad shallow paleo-channels are present on the higher elevations of southern New Jersey, especially in the headwaters of the Great Egg Harbor and Mullica rivers. These are difficult to recognize in the field but show clearly on the early 1930-1932 air photograph coverage of the area. One photograph, illustrated in Figure 5, shows part of the Great Egg Harbor and Penny Pot drainage systems north of Weymouth, Atlantic County. The present drainage is clearly misfit. The 'beaded' nature of the Penny Pot drainage pattern is especially puzzling. A striking aspect of many photos is the presence of braided paleo-channels. S. C. Farrell et al (1985) discuss similar channels on the neighbouring Mullica drainage system.

In modern fluvial systems, channel braiding and meandering are usually interpreted in the context of the amount and character of available sediment, the quantity and variability of discharge, and the mechanics of flow (e.g. Leopold, Wolman and Miller, 1964). We interpret the morphology of the palaeo-channels of the Pine Barrens uplands to reflect high seasonal (i.e. snowmelt-derived) flow over seasonally or perennially frozen ground combined with high sediment transport. Somewhat analogous fluvial conditions can be observed today on the tundra barrens or polar semi-desert terrain of Banks, Prince Patrick and other islands in the western Canadian Arctic that is underlain by Late-Tertiary-age Beaufort Formation sand and gravel.

Figure 5. Aerial photograph of part of northern Atlantic County, southern New Jersey, flown for the State of New Jersey, 1930-1932. The photo shows the paleo-channels of the Great Egg Harbor River and associated Penny Pot and Hospitality Branch streams in the area north of Weymouth, Atlantic County.

Rates of change

S. D. Stanford et al (2002) have investigated long-term rates and patterns of landscape evolution in north-central New Jersey. They conclude that Quaternary landscape incision and down-wasting occurred at a gross average rate of approximately 4mm/1000 years. This appears to be a relatively slow rate of landscape modification (see Goudie, 1995, 101-111). The widespread occurrence of low-angled pediment-like surfaces, mantled with Quaternary debris, suggests these surfaces have resulted from the headward retreat of upper slopes. As a result, landscape elements of varying antiquity are preserved.

Our own investigations in southern New Jersey support this interpretation (see below; stops 2, 4). Older surfaces are invariably those at higher elevations. Many appear to have experienced little change for many thousands of years.

The landscape evolution model proposed by S. D. Stanford et al (2002) is usually associated with hot arid regions. However, it is also applicable to cold arid regions such as Antarctica, interior Yukon and Alaska, and the never-glaciated parts of Siberia. Field studies suggest that extremely slow rates of landscape evolution prevail in such regions (e.g. French and Harry, 1992). In Antarctica, certain terrains are thought to have experienced little or no landscape modification for several millions of year (Summerfield et al, 1999).

In southern New Jersey, the substrates are unconsolidated, relatively young rocks, and the landscape elements accordingly much younger. It is reasonable to assume that enhanced landscape evolution would have commenced during the Early Pleistocene at the beginning and end of each cold period. These would have been times when sufficient moisture was available for both cold-climate mass wasting to occur on slopes and for surface flow to occur in channels. Rates of channel incision and amounts of sediment transport (i.e. denudation), would have varied temporally throughout the Quaternary in accordance with climatic fluctuations. The paleochannels of the Pine Barrens probably reflect the latest period of enhanced fluvial activity, towards the end of Late-Wisconsinan time, approximately 15-18 ka years ago. The last 10 ka years (i.e. the Holocene) must be regarded as a period in which the rate of landscape evolution in southern New Jersey has progressively decreased.

EXCURSION STOPS

STOP 1. VINCENTOWN – WOLFE'S SPUNG "THERMOKARST BASINS"

Topographic map: Mount Holly Quadrangle, N.J. Scale: 1: 24,000.
Latitude 39° 56'N; longitude 74° 45'W.
Road distance from Hammonton: ~22 miles.
Approximate time from Hammonton: 30-40 minutes.
Keywords: 'freeze-thaw basin'; 'spung'; deflation hollow; the beginning of the story

Almost 50 years ago, P. E. Wolfe drew attention to the numerous enclosed depressions and shallow basins that exists in southern New Jersey (Wolfe, 1953, 1956). He interpreted these as Late-Pleistocene 'freeze-thaw' phenomena, later referring to them as 'thermokarst basins' (Wolfe, 1977, pp 290-292). This stop allows us to view the depression that Wolfe described as his type-example.

'Freeze-thaw' basins

The basins in New Jersey are irregular in distribution; in some areas they occur in clusters, in others as isolated depressions or hollows. Some appear aligned. Most are irregular in outline and without apparent orientation. Some are oval or elliptical. Wolfe's basin is comparatively striking when compared to the multitude of similar depressions that exist elsewhere on the inner and outer coastal plains of New Jersey. It is also located in a relatively high topographic position on a river terrace. The underlying Vincentown Formation contains the only true marl found in New Jersey. The fine sandy loam subsoil (Tinton Sand series) is good for molding sand. Extensive quantities of marl were extracted from this general area of New Jersey during the 1920-1940 period for use elsewhere, especially in improving the sandy soils of the Pine Barrens.

The basins and depressions are of some antiquity. Water certainly existed in them over 12,000 years ago because archaeological evidence indicates that paleoindian sites were associated with many depressions. Many of these hunting camps were preferentially located on the south and west sides of the basins where there is a build-up of fine, sandy, wind-blown material (figure 6). Later, certain ponds ('spungs') were utilized as watering holes by some of the 17th and 18th century colonial trails that traversed the New Jersey Coastal Lowlands.

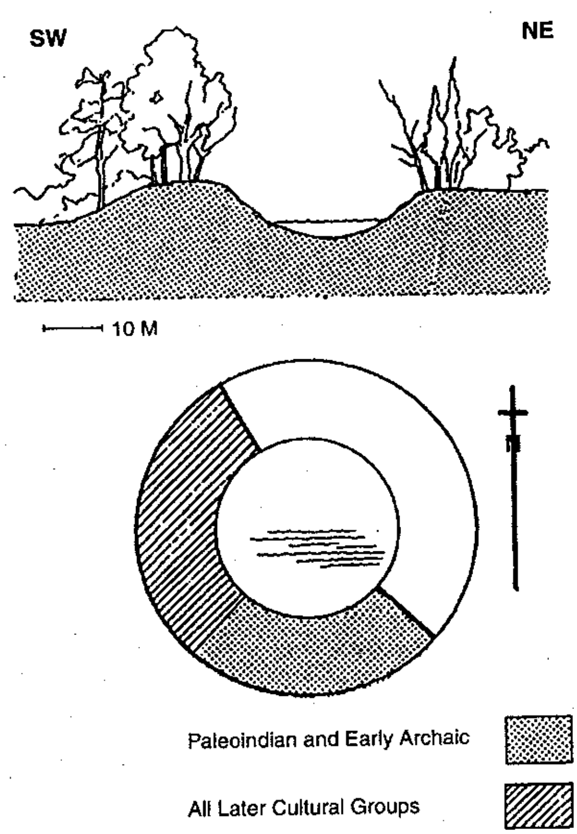

Figure 6. Diagram to illustrate schematic cross-section (not to scale) and the typical distributional relationships of cultural materials found on the raised rims of enclosed depressions (so-called 'periglacial basins'or 'pingo scars') in southern New Jersey. From Bonfiglio and Cresson, 1982, figure 6, p 35.

Origin

Wolfe's 'freeze-thaw' hypothesis was vaguely worded and his explanation was regarded as unconvincing (e.g. Rasmussen, 1953; Péwé, 1973, p 85). In subsequent years, the freeze-thaw idea was redefined when archaeologists referred to the depressions and ponds as 'pingo scars' (e.g. Bonfiglio and Cresson, 1982). It was observed that many depressions were enclosed by shallow ramparts, a feature they assumed diagnostic of a degraded frost mound (see French, 1996, 101-8, 249-253)

H. M. French and M. Demitroff (2001, 344-47) consider the various hypotheses that might explain the origin of these depressions and basins. These are:

1) Non-periglacial: (i) solution subsidence; (ii) water-upwelling and subsidence.

2) Periglacial (cold-climate): (i) Pingo-scar hypothesis; (ii) thermokarst-lake-basin hypothesis; (iii) erosion by nival processes; (iv) erosion by periglacial wind action.

Because carbonate rocks are absent, an interpretation involving karst solution and collapse is unlikely. Scaling problems and the absence of soft-sediment deformations in underlying Bridgeton and other units preclude water-upwelling and subsidence. The pingo-scar hypothesis is rejected for 3 reasons: (1) the topographic and hydrological conditions for either hydraulic (open) or hydrostatic (closed) system pingo growth are lacking, (2) water depths were insufficient for unfrozen zones (taliks) to have existed beneath the ponds, and (3) the number of enclosed depressions far exceeds the highest densities of pingos known to exist globally. The thermokarst-lake-basin hypothesis is rejected because sections excavated across several depressions fail to reveal thermokarst-lake sediments. The nival (snowmelt-runoff) hypothesis is unlikely because runoff would have been over impermeable (frozen) substrate, and the dimensions of many depressions indicate the volume of sediment that would need to have been removed downwards via 'sinks' is considerable.

By default, periglacial wind action is thought to be the cause of the enclosed depressions. They formed as 'blow-outs' in an area characterized by sparse vegetation at a time when strong winds dominated the landscape. The ramparts, best developed on southern and western rims by wind-blown sediment, suggest winds were from the north and east. As both sea level and the groundwater table rose in Late-Wisconsinan time, the depressions intersected the groundwater table and became wetlands. Ponds, recharged by groundwater, formed in some of the deeper basins. By 12-13,000 years ago Paleoindian hunting camps were located adjacent to the ponds. It is tempting to think that the south, west and southeastern rims were chosen because they were downwind of the water body, thereby minimizing the impact of wind-blown sediment upon daily life. Possible stages of evolution of the enclosed basins and wetlands of southern New Jersey are illustrated in figure 7.

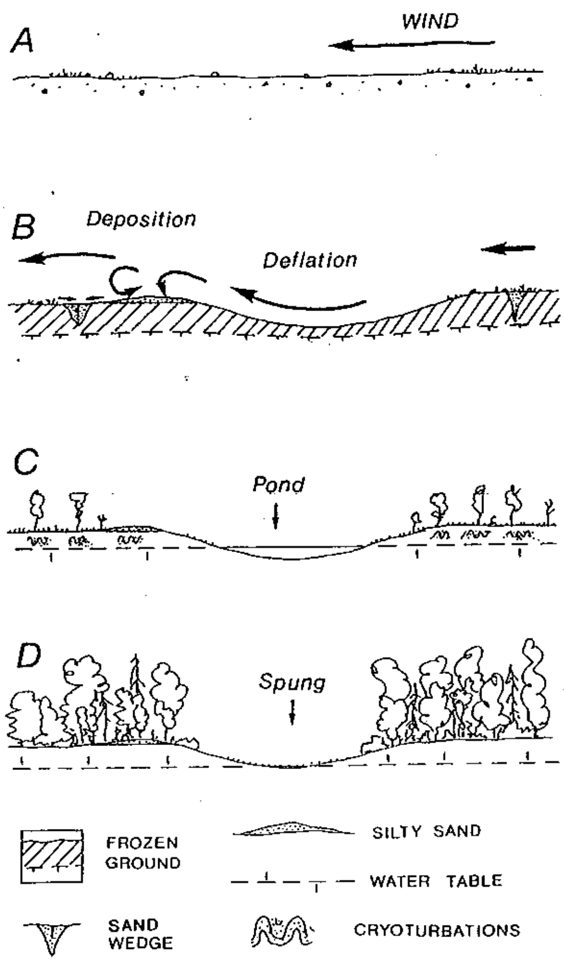

Figure 7. Possible stages in evolution of the enclosed basins and wetlands of southern New Jersey. Modified from French and Demitroff, 2001, figure 5, p 347.

A. Original Pine Barrens surface in Late-Pleistocene times.

B. During the maximum extent of Lat-Wisconsinan glaciation (~18-22 ka), perennially frozen ground (permafrost) and semi-desert conditions existed. Aeolian activity on exposed surfaces initiated deflation hollows or blow-outs. Thermal-contraction cracking occurred and sand wedges formed.

C. Deflation activity probably ceased by ~14-16 ka when an open woodland - shrub tundra vegetation developed. Permafrost degraded and thermokarst involutions formed. The regional groundwater table rose and depressions became wetlands and ponds, some utilized by Paleo-indian cultural groups as early as 12,500 years ago.

D. A gradual lowering of the regional water table in the 20th century has led to progressive shrinkage and disappearance of many ponds and wetlands.

STOP 2. PAWS FARM NATURE CENTER, MT — 'TIME TRAVELLERS' AND EVIDENCE FOR PLEISTOCENE WIND ACTION

Topographic map: Mount Holly Quadrangle, N.J. Scale 1:24,000.
Latitude 39° 58'N; longitude 74° 51'W
Road distance from Hammonton: ~30 miles
Road distance from Stop #1 to Stop # 2: ~8 miles
Approximate time between Stop # 1 and Stop # 2: ~15-20 minutes

Keywords: silcrete, ferricrete, ironstone, wind action, ventifacts

Background to this stop:

Wind is a well-known erosive agent. Wind-blown, rolling, and/or saltating mineral particles can effect abrasion upon many rocks and buildings. Sand particles (SiO^2) are commonly used to clean ancient monuments and other structures ('sand-blasting'). Ice is an equally effective abrasive agent; for instance, the Moh hardness of an ice crystal at -40° Celsius is 6.0, equal to that of steel.

'Silcrete' refers to a siliceous, indurated sand formed within a shallow subsurface as a result of weathering of bedrock or surficial deposits and later cemented at low temperatures and pressures. Silcrete is best known from low latitudes (e.g. Summerfield, 1983). There is debate as to its origin and exact mode of formation. It probably formed during subtropical to warm-temperate climatic conditions (Wyckoff and Newall, 1992). Other terms that are sometimes used include 'duricrust', 'sarsen', 'meulièr', and 'greywether'.

'Ferricrete' is another type off indurated rock, composed mainly of hematite, Fe_2O_3, but also associated with silica and hydrated sesquioxides (Fairbridge, 1968, 553-554). The formation of ferricrete involves complex geochemistry. Like silcrete, it may be the product of a warm environment.

PAWS Farm Nature Center

At PAWS, you have the opportunity to examine silcrete and related boulders. The silcrete is especially hard and resistant. The site is a small knob (hill), probably an outlier of Miocene-age rocks similar to those that form nearby Mount Laurel and Mount Holly. The hill is mantled with Quaternary-age colluvium. Elevation of the site varies between ~10-15 m above sea level. Although some of the larger stones have probably not been moved by man, equally, they cannot be regarded as being in-situ. Almost certainly, they have all moved downslope, under gravity processes.

New Jersey geological 'time-travellers'

As a generalization, silcrete occurs along the dissected scarp that separates the Delaware River Quaternary terraces from the upland surface underlain by Miocene sediments that constitutes the east-west watershed divide of southern New Jersey. The largest number of silcrete boulders occurs in the vicinity of Swedesboro where they litter gentle slopes and flat surfaces 15-20 m above sea level. (Wyckloff and Newall, 1992). Elsewhere on the higher elevations of southern New Jersey, scattered blocks of orthoquartzite (locally known as 'cuesta quartzite') are also found at approximately the same elevations as at Swedesboro.

The silcrete and orthoquartzite boulders are probably the remnants of a duricrust that formed in a 'feather-edge' situation associated with the outcrop pattern of certain sand and clay members of the Kirkwood Formation (Wycloff and Newall, 1992, 42). The silcrete is thought to be Miocene in age with cementation in early Pliocene time. Being extremely resistant, the silcrete has survived through the Quaternary. Thus, the boulders are geological 'time-travellers'.

Ferricrete blocks also occur in southern New Jersey in the same general distributional pattern as the silcrete. They appear related to the bog ironstone that occurs widely in the area. Today, bog ironstone forms around the edges of wetlands and along stream banks. By its very nature, bog ironstone forms at the level of the regional water-table. In all probability, ironstone formed continuously throughout the Quaternary, and perhaps earlier, in accordance with fluctuations of the groundwater table. Probably, the ferricrete developed in an analogous fashion to silcrete at localities where bog ironstone was exposed on a palaeo-ground-surface. We must assume that the ferricrete is also pre-Quaternary in age for induration to have occurred and that, like the silcrete, it is a geological 'time-traveller'.

Judging from the local topography, the larger silcrete boulders have moved a maximum of ~350 m, as inferred from the nearest regional interfluve, and as little as ~ 50 m, if the PAWS cemetery is judged to be the nearest local interfluve. Thus, the inferred distances moved by the New Jersey time-travellers are considerably less than the 1000-1500 m reported for analogous (i.e. sarsen) boulder movements in southern England (e.g. Williams, 1968; Small et al, 1970). This probably reflects the lower relative relief of southern New Jersey. It may also indicate a lower rate of mass wasting in southern New Jersey.

Wind action and the geological 'time-travellers'

The larger silcrete boulders show clear signs of wind sculpting and polishing (figure 8A). It is clear that this abrasion is not occurring today. However, because all boulders have moved, either naturally or by man, it is impossible to infer paleo-wind direction. It is debatable whether the abrasive agent was wind-blown silt and fine sand, or snow crystals. What is clear, however, is that these ancient boulders show evidence of intense wind action and abrasion that is unrelated to today's climatic environment.

Numerous smaller silcrete boulders are now incorporated along the walkways and borders on the farm. Many are strongly ventifacted (i.e. show preferential abrasion and 'edging'). Both 'einkanter' (one-sided) and 'driekanter' (two-sided) forms can be found. They suggest abrasion was caused by the saltation and transport in suspension of erosive particles near the ground surface. By contrast, the larger silcrete boulders that occur in the trees along the Nature Trail show evidence of scalloping, helical score and polishing by high wind velocities at some elevation above the ground surface.

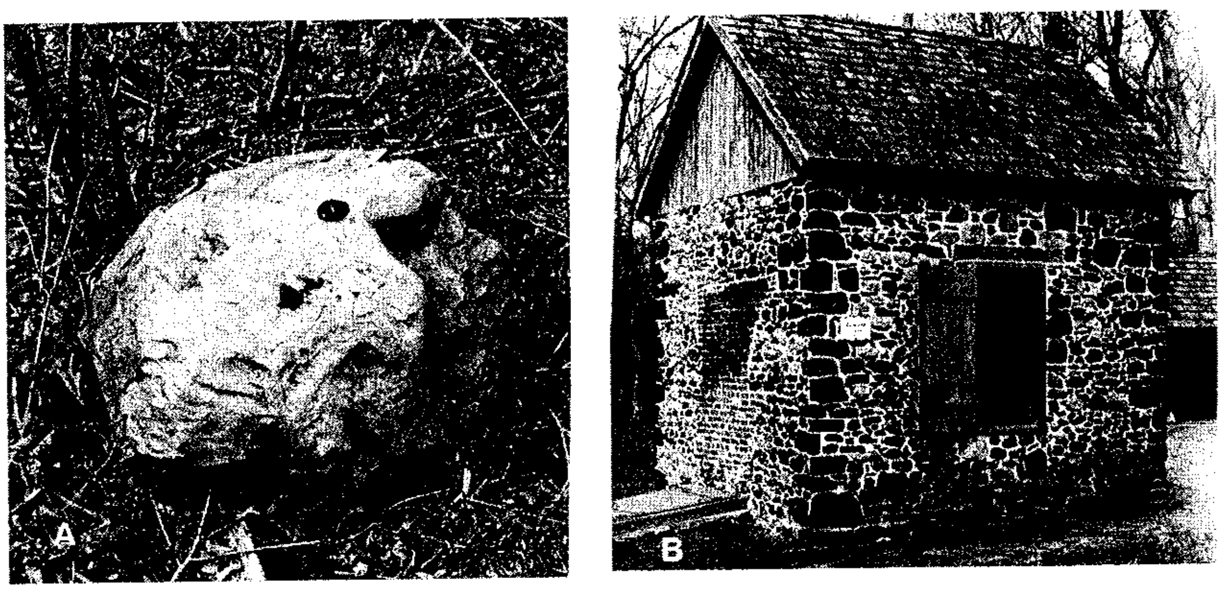

Figure 8. PAWS Farm Nature Center.

A. large silcrete ('cuesta quartzite') boulder showing wind-scalloping and polishing;

B. The 'icehouse', initially dating from 1700 but having been rebuilt using blocks of local silcrete, ferricrete and bog ironstone. Both photos were taken in April 2003.

Several questions remain unanswered at PAWS:

1) What were the magnitudes, directions and frequencies of the winds involved…?
2) What were the abrasive agents involved in the wind-modification of these extremely old and resistant boulders…?
3) How far have these boulders moved from their initial place of formation…?
4) Was movement by traditional creep or by solifluction…?

Human response

Sicrete makes fine building material if it is dressed. A number of these local-indurated rocks have been incorporated into the walls of the recently-restored 'smokehouse' and 'icehouse', initially built in 1700 and 1730 respectively (figure 8B). The Ice House is comprised of blocks of ironstone (~20%), ferricrete (~50%) and silcrete ('cuesta quartzite') (~30%). The ferricrete blocks, being more suited to angularity, have been preferentially used for the corners of the building. The contrasts in mineralogy are clear.

Elsewhere in southern New Jersey, many of the larger silcrete, orthoquartzite and ferricrete boulders were quickly collected by the early settlers and used for buildings and construction (e.g. Bonnie's Bridge, south of Route 38 in Cherry Hill); others have been collected more recently by local residents and are now employed as decorative items in gardens. Unfortunately, therefore, few boulders are in-situ, and even fewer are capable of revealing the intimate details of their travel through time.

STOP 3: PAKIM POND, LEBANON STATE FOREST- 'THE PINE BARRENS'

Map: Brown Mills Quadrangle, New Jersey, scale:1: 24,000.
Latitude 39° 52'N; longitude 74° 31'W.
Road distance from Hammonton: ~51 miles
Road distance between stop # 2 and stop #3: ~21miles
Approximate travel time between stop # 2 and stop #3: ~25 minutes

Keywords: The Pine Barrens; spungs; human occupance; lunch

Pakim Pond represents a typical Pine Barrens environment. This is also the LUNCH STOP – Southern Cookin'!

Human occupance in the Pine Barrens

Archaeological evidence indicates that Paleo-Indians were moving through the Pine Barrens 12,000 – 8,000 years ago (Mounier, 2003). Archaic (8,000 BC – 3,000 BC) and then Woodland Indian (3,000 BC – 500 AD) camps followed. Finding sufficient lithic materials for tool making was a problem for these aboriginal populations who crossed the sandy terrain. The archaeologist J. A. Cresson has documented the significance of prehistoric quarrying and working of silcrete by the native populations of South Jersey.

The first European settlement was in the middle of the 17th century, particularly after the English seized New Jersey from the Dutch in 1664. Bog ironstone, currently forming along the river banks and at the edges of ponds, and local charcoal became the basis for a flourishing iron smelting industry in the 18th and 19th centuries. Glass making, using local sand, was also important. Following decline of the iron industry, and over-exploitation of the forest resources, the economy of the Pine Barrens declined. The railways opened up settlement possibilities further west. However, demands for lumber and charcoal continued and from 1750 until 1930 the Pine Barrens were largely devoid of mature trees. The majority of the present Pine Barrens forest is re-growth since the 1930's. Since the late 19th century, the use of wetlands for cranberry, and later blueberry, cultivation is a trend that continues to this day.

The Pine Barrens supports a rich local culture. The legend of the 'Jersey Devil' was spawned in the area in the 17th century. Today, some of the deeper ponds have folklore associated with them (e.g. the 'Blue Hole' near Winslow). Many of the old colonial trails traversed the Pine Barrens and place names reveal the origins of past settlers, and sites of previous economic activity (e.g. Hampton Furnace; Bennetts Mill). Certain settlements no longer exist and are 'ghost' towns (e.g. Carmantown); others are shadows of their former past. There are rumours of a 'lost silver mine'. In the 19th century, the area was part of the black 'underground railroad'. In the 20th century, the area continued to be settled by Italian, Ukrainian, German, Jewish, Puerto Rican and other migrant groups. For many years illicit stills operated in the more remote and inaccessible areas. We say no more...

In 1978, the US Congress established the Pinelands National Reserve and in 1979 the New Jersey State Legislature enacted the Pinelands Protection Act and thereby created the Pinelands Commission. The role of the Pinelands Commission is to preserve, protect and enhance the natural and cultural resources of the Pinelands National Reserve, and to encourage compatible economic and other human activities consistent with that purpose. The Commission monitors development within the Pinelands, acquisition of land, planning, research and education.

BUS TRANSECT THROUGH THE PINE BARRENS

Distance 40 miles.
Approximate travel time: 60-70 minutes
Route: Chatsworth - Jenkins Neck- Green Bank - Lower Bank - Egg Harbor - Mays Landing - Dorothy - Hesstown - Manumuskin - Port Elizabeth.

STOP 4: DORCHESTER PIT NEAR PORT ELIZABETH – ANCIENT SAND WEDGES: EVIDENCE FOR LATE-PLEISTOCENE PERMAFROST

Topographic map: Port Elizabeth Quadrangle, New Jersey, scale 1:24,000
Latitude 39° 17' N; longitude 74° 57'W.
Distance from Hammonton: ~107 miles
Road distance between stop # 3 and stop # 4: ~40 miles
Approximate travel time between stop # 3 and stop # 4: ~60-70 minutes.

Keywords: Late-Pleistocene sand wedges; chronostratigraphy; Bridgeton Formation; luminescence and C^{14} dating

The objective of the stop is to demonstrate that permafrost conditions existed in southern New Jersey during the Late-Pleistocene. A more detailed discussion is provided in French, Demitroff and Forman (2003).

Geomorphological setting (figure 9)

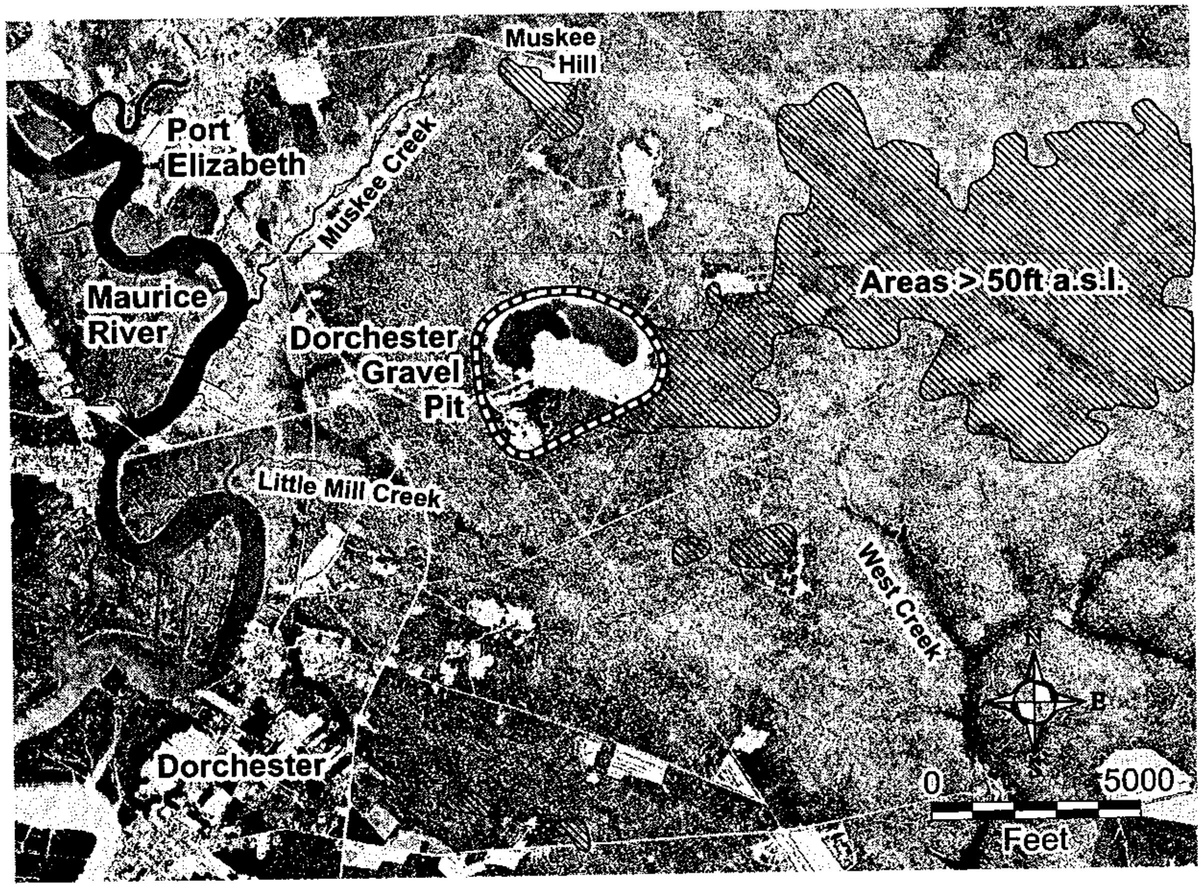

Figure 9. A United States Geological Survey aerial photograph (1995/97) showing the location of the Dorchester Pit, southern New Jersey, with main geomorphological and locational features indicated.

This large commercial pit is located on a flat surface that lies at an elevation of ~15-18 m above sea level. This relatively well-drained upland surface extends north-eastwards from the Maurice River. The surface is dissected to the north by Muskee Creek, a small tributary draining to the Maurice River at Bricksboro. Muskee Hill, at an elevation of 20m (67') above sea level is an outlier of this surface. Further remnants of the surface occur to the west of Dorchester but here the surface has been extensively dissected by Little Mill Creek and Crowder Run.

The pit is excavating material from the Bridgeton Formation (Tbr). This unit occurs widely throughout the central parts of southern New Jersey. It consists of a highly-weathered mix of sand, silt, clay, cobbles and boulders. Clay mineralogy best illustrates the long period of leaching and subaerial oxidation that these sediments have experienced (figure 10). The Aura soils (Tedrow, 1986, 310-310; 'typic fragiudult', Soil Survey Staff, 1999) dominate the yellowish-red sands and gravels of the Bridgeton Formation. They occur over wide areas of Cumberland, Salem, and Gloucester counties. Of special interest are (a) a fragipan (indurated) layer in the B/C soil horizon, (b) a soil matrix that is strikingly vesicular in nature, and (c) an A horizon that is partially aeolian in origin. All these soil characteristics can be seen in the sections exposed in this pit. We suggest all have cryogenic significance.

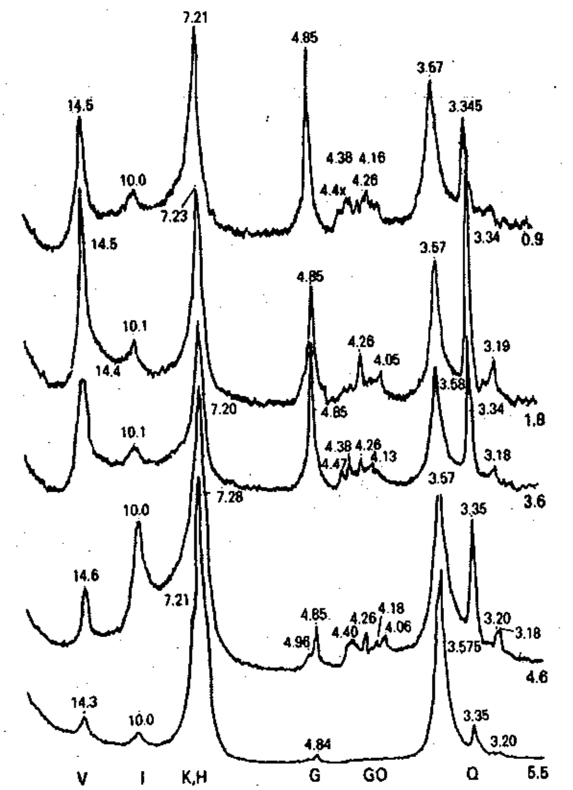

Figure 10. X-ray-diffraction graphs of the clay-silt fraction obtained from near-surface gravels of the Bridgeton Formation (Tbr) near Hammonton, N.J. Sample depths are indicated in meters. Measurements are in angstroms. Gibbsite (G) and vermiculite (V) increase in abundance from depth towards the surface. Kaolinite (K, H) decreases in the same interval. GO is goethite and Q is quartz. The clay mineral assemblage is mature. From Owens et al, 1983, figure 28, p. F35.

Sand of the underlying Cohansey Formation forms the upper slopes surrounding the gravel surface. Various Quaternary-age alluvial and estuarine deposits (Qs, Qcm_1, Qcm_2, Qcm_3) cover the lower slopes, valley bottoms and riverine lowlands (Newell et al, 2000).

Ancient sand wedges

Two sets of Late-Pleistocene-age sand wedges have been identified in this pit. Their chronostratigraphy is illustrated in figure 11.

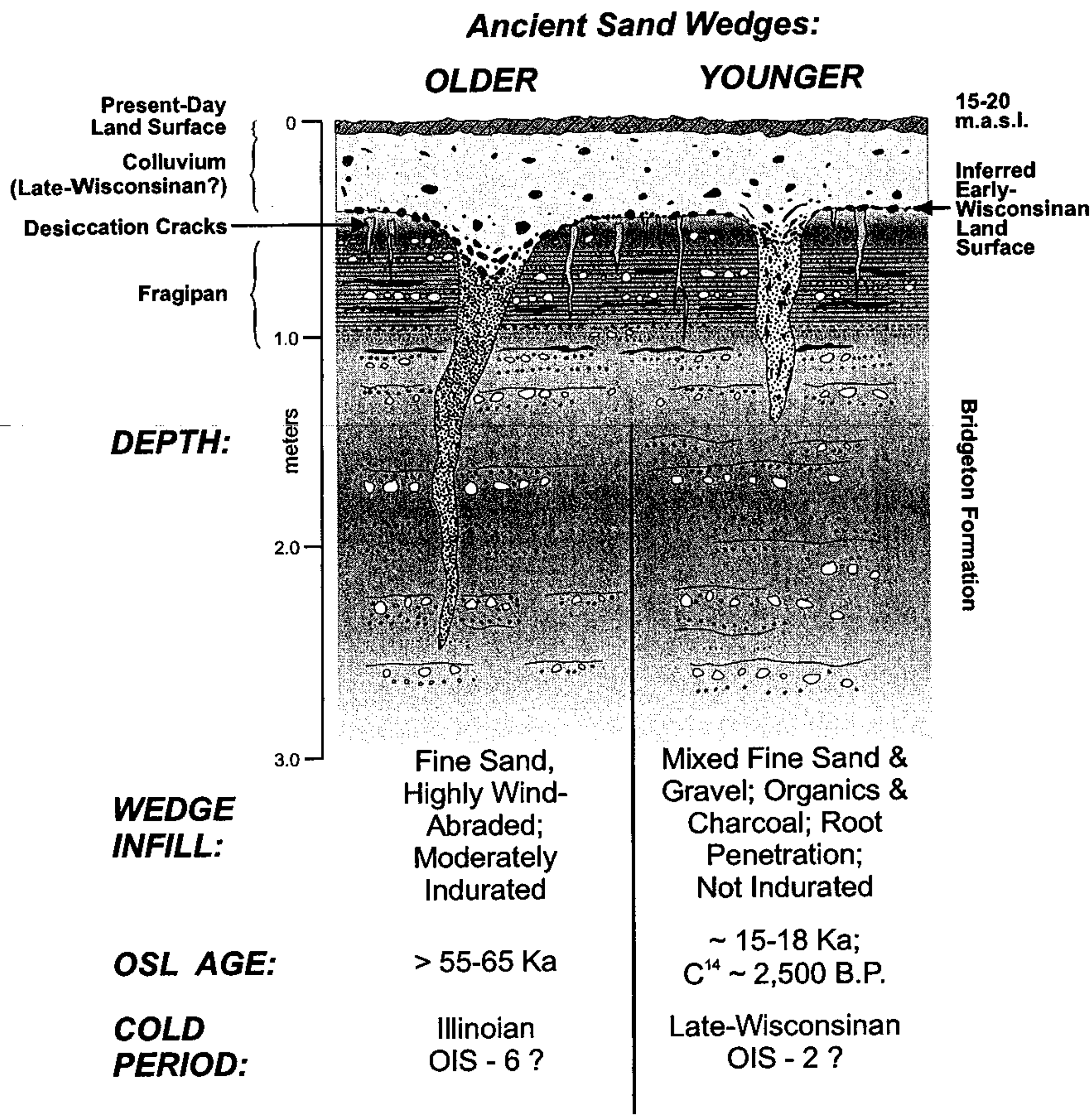

Figure 11. Schematic diagram illustrating the Late-Pleistocene cold-climate chronostratigraphy of the Dorchester Pit, southern New Jersey.

The wedges formed through thermal-contraction-cracking of the ground. As such, they indicate the previous existence of permafrost. The cracks have been filled with a combination of silty sand and locally-derived sand and gravel particles. The wedges are between 1.0-2.5 m deep and 0.2-0.4m wide and all occur within 0.5-1.0 m of the ground surface.

One set (figure 12A)) is filled with highly wind-abraded fine sand particles and typically illustrates vertical or near-vertical laminations. The other (figure 12B)) contains loosely-packed, non-oxidized sandy gravel. The latter also contain rootlets and fauna associated with the present forest cover that exploit the lines of weakness presented by the wedges. Both types of wedge are relatively easy to identify when exposed in section because the infill material contrasts with the indurated, stained and highly-weathered nature of the enclosing Bridgeton Formation gravels.

Figure 12. Examples of Late-Pleistocene sand-wedge structures, Dorchester Pit.

A. A wedge, ~2.7m deep and 5-20 cm wide, containing well-sorted and highly wind-abraded sand. The quartz fraction is dated at >65ka.

B. A wedge, ~1.7 m deep and 25-30 cm wide, containing loose sandy and pebbly gravel. There is abundant root penetration. Both quartz and polymineral fine fractions are dated at ~15 ka. A C^{14} age from root fragments gave an age of 2580 years BP. Both photos were taken in August 2002.

The wedges are not regularly spaced, as might be expected if they formed a polygonal net, and do not appear to have any visible surface expression. Because all the wedges appear related to the present ground surface, they must post-date the age of that surface.

In several ways, the wedge structures are unlike the typical ice-wedge pseudomorphs commonly described in the permafrost literature. First, the infill is of a primary nature; second, there is no downturning or 'sagging' of the adjacent enclosing sediments that would normally accompany the thaw-deformation of an ice wedge; third, the sand grains within the wedge are predominantly well sorted, well rounded and 'frosted', all characteristics of wind transport and abrasion (figure 13). The wedges are interpreted, therefore, as sand wedges, formed by the thermal-contraction-cracking of perennially frozen ground (permafrost).

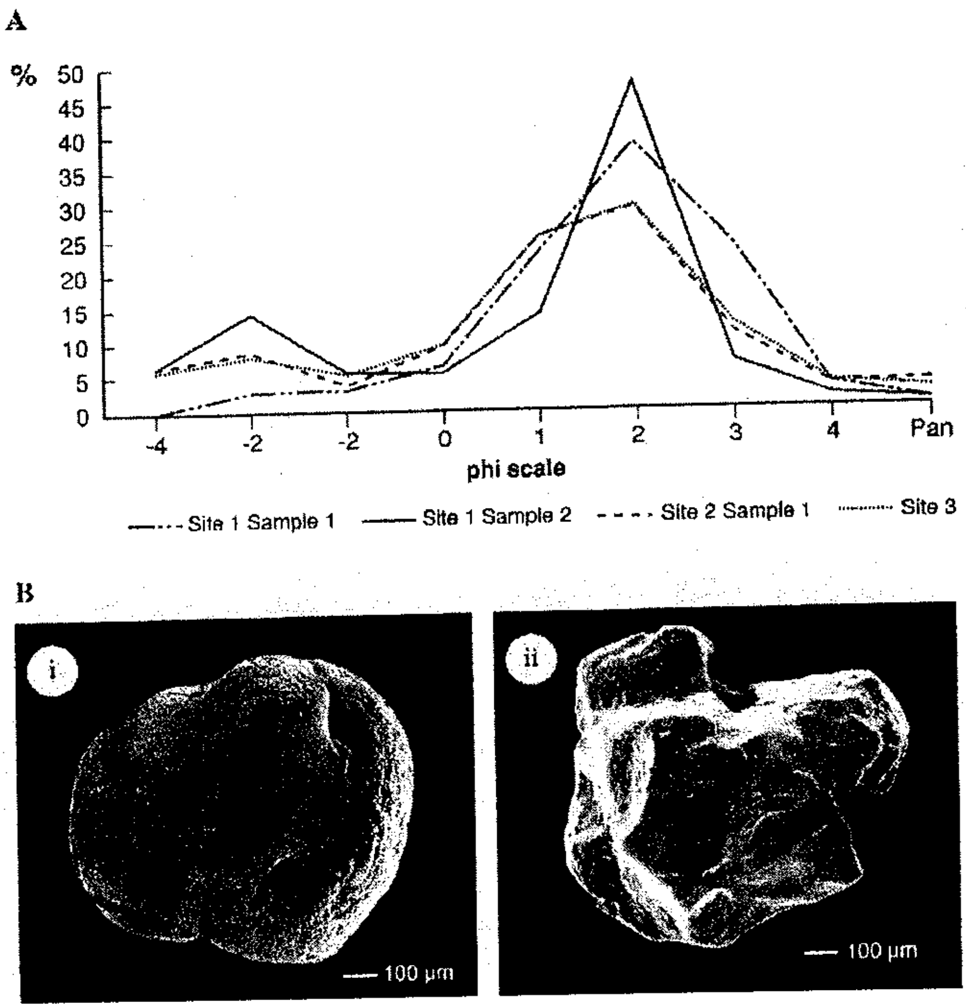

Figure 13. Typical granulometry and grain morphology of sand particles within the wedge structures, Dorchester Pit:
(A) Grain-size distribution of four sand samples taken from within 3 different wedges. The fine sand fraction constitutes between 65-83% of each sample analyzed.
(B) SEM photographs of (i) a wind-abraded sand grain from within a wedge and (ii) a non-wind-abraded sand grain from adjacent Bridgeton Formation (Tbr) gravels. Magnification: x100.

In terms of palaeoclimatic significance, thermal-contraction-cracking in coarse-grained substrates is generally thought to imply a maximum mean annual air temperature (MAAT) of <-4° C and > -8° C (to -6° C). The sparse tundra vegetation that would have existed on the sandy substrate of the Pine Barrens, combined with the relatively high thermal conductivity of such materials, would have meant that the 'thermal offset' would have been reduced. Therefore, a MAAT of between -3° and -4° C may be a more reasonable inference. The wind-abraded sand infill to the wedges indicates strong wind action and a relatively arid environment.

The lack of a visible polygonal pattern on the ground surface suggests that the conditions suitable for thermal-contraction cracking were marginal in the Pine Barrens. Analogous conditions today are probably represented by the incomplete and spatially discontinuous polygon patterns that exist in parts of the northern boreal forest of North America (e.g. south of Inuvik, NT, or in the Mayo region, YK). In these areas, wedges are thought to be either inactive or the result of local cold-air drainage

Age

Dating of the wedges has been undertaken using both radiocarbon (C^{14}) and infrared optically-stimulated luminescence (IRSL) methods. Radiocarbon dating is suited to organic materials. Luminescence dating subjects mineral grains, mainly quartz and feldspar, to radiation and takes advantage of the time elapsed since deposition of the grains.

Three samples of the coarse-grained (100-150μm) quartz fraction from within three different wedges yielded infinite values of >66 ka and >55 ka and a finite value of ~15 ka respectively. Less reliable dates obtained from the polymineral fractions of the two older samples gave ages in excess of >130 and 198 ka. Collectively, the OSL dates indicate that two sets of wedges of different ages are present at this site. The older dates correlate with an early Wisconsinan, or possibly older (Illinoian; oxygen-isotope-stage 6?) cold period. The younger date correlates with the Late-Wisconsinan maximum (oxygen-isotope-stage 2), ~15-22 ka years ago. A C^{14} age determination from charcoal fragments within the younger wedge gave a conventional age of 2580 +/-40 yr BP. This sample is interpreted as a rootlet that post-dates wedge formation and probably indicates a late-Holocene fire episode.

At both time periods when permafrost was present, the continental ice-sheets would have reached into north-central New Jersey. Because both sets of wedges are epigenetic rather than syngenetic in nature (i.e., the wedges are younger that the land surface on which they formed), considerable antiquity and stability of that surface is also implied. Our findings support, therefore, the conclusions of S. D. Stanford et al (2001) that "...surfaces, while active, were not significantly eroded under periglacial conditions but rather modified in place" (p. 271). They also support the inferences made by pedologists (e.g. Krebs and Tedrow, 1958) that the red-yellow podzolic soils of southern New Jersey formed during a period of landscape stability that probably lasted through one or more interglacials.

STOP 5 (Optional). A PIT NEAR NEWTONVILLE – CRYOTURBATIONS: EVIDENCE FOR LATE-PLEISTOCENE THERMOKARST

Topographic Map: Buena Quadrangle, New Jersey, scale 1:24,000
Location withheld by request of the pit operator. **PERMISSION REQUIRED FOR ACCESS.**

Keywords: Thermokarst involutions; cryoturbations;

This stop allows participants the opportunity to examine a number of involuted and disturbed sedimentary structures that are developed in loamy and sandy materials (figure 14). Such structures are commonly termed 'cryoturbations' (see French, 1996, 240-243). Similar structures have been described widely from near-surface sediments in both central and southern New Jersey. Typically, the deformations extend to depths as great as 2-4 m below the ground surface.

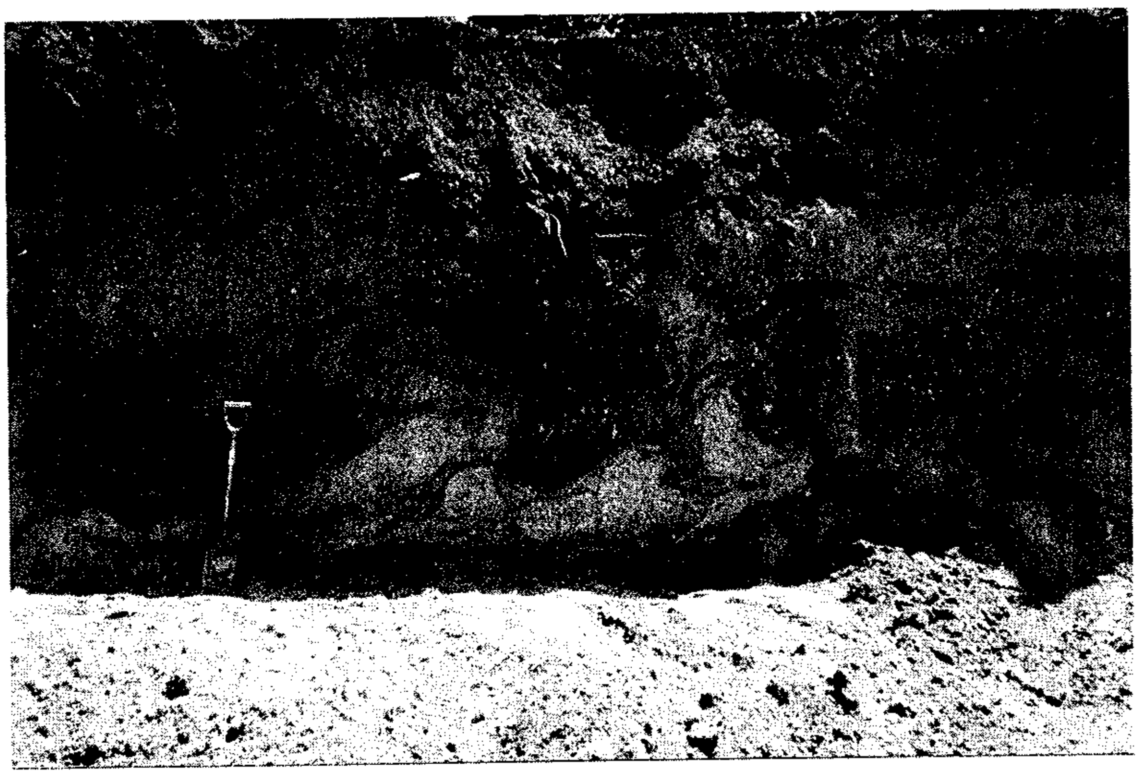

Figure 14. Late-Pleistocene cryoturbation structures exposed in a pit near Newtonville, Atlantic County. See text for description of these sediments. Photo was taken in May 2001.

Location

The site is a commercial sand and gravel operation. The area has produced aggregate products since the 1930's. It is a well-drained ridge or upland surface at an elevation of ~40 m above sea level This is one of the highest elevations in Atlantic County. The ridge is mapped (Newall et al, 2000) as being capped by gravels assigned to the Bridgeton

Formation. However, we interpret the sand exposed in the pit as being a lower Bridgeton unit that probably rests immediately upon sand of the Cohansey Formation.

The sand

The sand is highly variable in nature. It consists of layers and pockets of coarse, strongly bonded, reddish-brown and yellow molding sand, fine clayey sand, and clayey coarse sand. All are frequently deformed. In fact, the protean nature of these deposits was a source of constant frustration to the early operators of this pit (personal communication, Richard Brimfield, previous owner). The worthless clayey-sand residue eventually proved ideal for the inner cover of the local Brimfield ballfield.

The well-drained soils formed on these loamy sandy substrate are termed Sassafras-Hammonton phase (Downer) (Tedrow, 1986, 279-285; 'typic hapludult', Soil Survey Staff, 1999). They occur widely in Cumberland, Atlantic and Camden counties as well as further north in South Jersey. They are generally impoverished and acidic, and contain few minerals other than quartz. Tedrow (1986, 283) describes the soil as 'xeric-like or transitional from well-drained to excessively drained'. Unquestionably, extensive leaching has occurred but, because of the lack of feldspars and clay minerals, the soil lacks the prominent B/C (fragipan) horizon that characterises the Aura soil of the Bridgeton Gravels. Therefore, although a distinctive clay mineralogy pattern is absent, there is every reason to assume that these sands, like the Bridgeton Formation gravels, have been exposed to periods of extensive subaerial weathering.

The heterogeneity of the sand deposits at this locality may be the reason why cryoturbation structures are well developed. Under permafrost conditions, ice segregation would have caused ice lenses to preferentially form in the clayey layers.

Possible origins of the structures

The periglacial literature indicates two possible mechanisms:

1) They are soft-sediment deformations that are associated with the thaw of permafrost (i.e. 'thermokarst involutions'). In this case, density-controlled mass displacements that occurred during thaw of ice-rich, water-saturated sediments cause the clayey sands to ascend and the overlying coarser sands to descend.

2) They formed during the repeated freezing and thawing of a seasonally-frozen, near-surface layer. In this case, it is argued that freezing was from the surface downwards. The overlying rigid (i.e. frozen) carapace prevented upward heave of the more clayy layers and, instead, induced a progressive downward movement of material into unfrozen sediment below.

Interpretation

We interpret these structures as 'thermokarst involutions'. They formed during progressive permafrost thaw when especially icy layers within the permafrost melted (see Eissmann, 1978; Murton and French, 1993; Murton 1996, 2001). The main reason why explanation #2 is inappropriate is that the active layer in unconsolidated sediments in permafrost regions today rarely exceeds 1.5 m in thickness. This situation reflects not only the air temperature magnitude but also ground thermal considerations (e.g. the Stefan equation). The latter is commonly employed in permafrost engineering practice to predict depths of freezing and thawing. Even in Central Siberia today, where summer air temperatures can exceed 30° C, thaw depths in unconsolidated sediments in excess of 2.0 m are rarely attained. A second reason is that freezing-induced, or so-called 'cryostatic', pressures have yet to be convincingly measured in field situations.

The time when these structures formed is difficult to determine. By definition, periods of permafrost degradation, i.e. thermokarst, would have followed each period of permafrost aggradation. The nature and amount of deformation would have reflected the degree of consolidation of the sediment involved, the speed of thaw, and the amount of ground ice present within the thawing permafrost body.

ACNOWLEDGEMENTS

We wish to acknowledge funding from the Great Egg Harbor Watershed Association (GEHWA) via the US National Park Service, Wild and Scenic Rivers Program. We also wish to thank PAWS Farm Nature Center and Mays Landing Sand and Gravel for allowing access to their properties at Mt Laurel and Port Elizabeth respectively. Several other commercial pit operators and numerous local residents of the Pine Barrens have helped us in many ways. They are too numerous to mention individually but are collectively thanked here. Helpful comments on the manuscript were provided by Michael Hozik, Mark Mihalasky and Scott Stanford. Mark Mihàlasky also assisted in the preparation of some of the diagrams.

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A Glossary of Permafrost, Periglacial, Pleistocene and New Jersey Pinelands Terms

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NEW JERSEY PINELAND TERMS:

'Spung' – an old English term for a pocket or purse. The word is given to the enclosed depressions and basins of southern New Jersey that constitute wetlands. Spungs are closed systems that are internally drained.

'Cripple' – short, broad, and damp, but wooded, paleovalley, which lacks a modern incised stream channel. Unlike spungs, cripples are open systems that drain into larger waterways.

'Blue Hole' – a name given to a small, often circular, riverine spring or upwelling in the Pine Barrens. Blue holes are often between 2-4 m in depth. Some are reputed to be 'bottomless', others to possess whirlpools and/or 'devils'. Blue holes contain clear water in contrast to the tea-stained waters common to spungs and cripples in the Pinelands.

'Ironstone' – bog ironstone, the result of chemotrophic bacteria that oxidize soluble iron in the groundwater. Bog ironstone forms at the level of the regional water table. Natural organic substances combine with the iron to produce the dark tea-colored waters in spungs and cripples.

'Savannah' – an open, grassy wetland with scattered trees. It reflects an edaphic condition induced by a high water table and fire. It typically occurs on the broad paleochannels. Once numerous, savannahs have largely disappeared from the Pinelands landscape, probably because the regional water table has dropped.

'Plain' – a broad, high, flat surface with stunted growth that is the result of the well-drained topography and frequent fires. Plains were once scattered throughout the Pine Barrens but are now confined to the area least disturbed by human activity.

PERMAFROST TERMS:

Active layer – The top layer of ground subject to annual thawing and freezing in areas underlain by permafrost.

Aggradation (growth) of permafrost – the increase in thickness and/or extent of permafrost.

Cryofrost – the 0° Celsius temperature line in a freezing material.

Degradation (thaw) of permafrost – the decrease in the thickness and/or extent of permafrost. Often termed 'thermokarst'.

Depth of zero annual amplitude – the distance from the ground surface downwards to the level where there is practically no annual fluctuation in ground temperature

Freezing (of ground) – the change of phase from water to ice in either soil or rock.

Freezing front – the advancing boundary between frozen (or partially unfrozen) ground and unfrozen ground.

Frost-stable ground – ground (soil or rock) in which little or no segregated ice forms during seasonal freezing.

Frost-susceptible ground – ground (soil or rock) in which segregated ice will form under the required conditions of moisture supply and temperature.

Frozen fringe – the zone in a freezing frost-susceptible soil between the warmest isotherm at which ice exists in pores and the isotherm at which the warmest ice lens is growing.

Frozen ground – soil or rock in which part or all of the pore water consists of ice.

Geothermal gradient – the rate of temperature change with depth in the ground.

Ground thermal regime – a general term encompassing the temperature distribution and heat flows in the ground and their time-dependence.

Ground ice – a general term referring to all types of ice forming in freezing and frozen ground

Permafrost – ground (soil or rock) that remains at or below 0°C for at least two years. Spatially, it is divided into zones of continuous, widespread discontinuous and scattered discontinuous permafrost where respectively over 80%, 30-80%, and less than 30% of the terrain is underlain by perennially frozen ground.

Permafrost base – the lower boundary of permafrost.

Permafrost table – the upper boundary (top) of permafrost (i.e, the bottom of the active layer).

Seasonally-frozen ground – ground that freezes annually.

Seasonally-thawed ground – ground that thaws annually.

Talik – a layer or body of unfrozen ground in a permafrost area. It is often an aquifer. By contrast, permafrost is often impermeable and acts as an aquiclude.

Thermal offset – the temperature difference between the mean annual ground surface temperature (MAGST) and the temperature at the top of permafrost (TTOP). The magnitude of the thermal offset depends largely upon the thermal conductivity of the earth materials in question and the nature of the vegetation cover.

Thermokarst – the process by which characteristic landforms result from the thaw of ice-rich permafrost

PLEISTOCENE AND RELATED TERMS:

The Quaternary Era -- the last 2-3 million years (i.e. the period of geological time in which there is evidence for the existence of homo-sapiens and predecessors.

'The Ice Ages' – a popular term for the Quaternary Era, referring specifically to the colder periods when continental ice sheets developed in the high latitudes.

The Holocene – usually refers to the last 10,000 years.

The Pleistocene – that part of the Quaternary Era that is not part of the Holocene.

The Late-Pleistocene – usually refers to the last ~ 250,000 years.

The Wisconsinan – the last major cold period, ~70,000-12,000 years ago. The Early-Wisconsin Glacial Maximum lasted from ~50,000 to ~70,000 years ago. The Last Glacial Maximum (LGM), termed the Late-Wisconsinan, occurred ~18,000-24,000 years ago. At the LGM, the continental ice-sheet reached as far south as northern New Jersey, New York City and Long Island.

The Illinoian – the penultimate major cold period, probably lasting between ~120,000-160,000 years ago.

The Sangamon – a period of warmer, more temperate conditions that existed between the last major cold period (Wisconsinan) and the penultimate major cold period (Illinoian), i.e. ~70,000-120,000 years ago.

Pre-Illinoian glaciations – a number of major cold periods, of varying intensity and duration, are thought to have characterized the Middle- and Early-Pleistocene.

Oxygen isotope stages (OIS) – Oxygen isotopes values, derived mainly from ice cores and marine sediments, together with corals and sea level fluctuations, enable a global chronostratigraphy to be developed for the last 1-2 million years. Cold glacial stages are assigned even numbers and warm interglacial stages are assigned odd numbers. The Holocene is OIS-1; the Late-Wisconsinan cold period is OIS-2; the Illinoian cold period is OIS-6.

Radiocarbon dating – Dating using the residual concentration of C-14 left in organic material.

Luminescence dating – Dating using the time elapsed since deposition of mineral grains, mainly quartz and feldspar, and then subject to radiation.

PERIGLACIAL AND RELATED TERMS:

Cryoturbations – irregular structures formed in earth materials by deep frost penetration and frost action processes, and characterized by folded, broken and dislocated beds and lenses of unconsolidated deposits, including organic horizons and even bedrock.

Fragipan – an indurated zone formed by clays and sesquioxides that have accumulated in the B/C horizon through weathering and leaching, usually over a long period of time.

Frost action ('freeze-thaw') – the process of alternate freezing and thawing of moisture in soil, rock or other materials, and the resulting effects on materials and structures placed on, or in, the ground.

Ice wedge – A massive, generally wedge-shaped body with its apex pointing downwards, composed of foliated or vertically banded, commonly white, ice.

Ice-wedge cast – a filling of sediment in the space formerly occupied by an ice wedge

Mass wasting – the downslope movement of soil or rock on, or near, the earth's surface under the influence of gravity.

Periglacial – the conditions, processes and landforms associated with cold, nonglacial environments.

Pleistocene 'periglacial zone' – the ice-marginal area adjacent to the continental ice sheets that were characterized by intense cold, strong winds, and tundra-like conditions.

Pingo – a perennial frost mound consisting of a core of massive ice, produced primarily by injection of water, and covered with soil and vegetation.

Pingo scar – a pingo remnant in a contemporary, non-permafrost environment

Polygons – Macro-scale polygons, typically 15-30 m across, result from the thermal-contraction-cracking of the ground in permafrost areas. The cracks form random orthogonal or oriented patterns at the surface.

Sand wedge – a wedge-shaped body of sand produced by the filling of athermal contraction crack with sand either blown in from above or washed down the walls of the crack.

Sand-wedge remnant (ancient sand wedge) – A sand wedge in a contemporary, non-permafrost environment.

Solifluction – the slow downslope flow of saturated unfrozen earth material. The term is commonly applied to processes operating in both seasonal frost and permafrost areas. In permafrost areas, the term gelifluction is sometimes used.

Thermal contraction cracking – the process by which tensile fractures result from thermal stress in frozen ground. In permafrost regions today, thermal-contraction-cracking usually requires a MAAT of -4°C or colder.

Thermokarst lake (depression) – a lake (area) occupying a closed depression formed by settlement of the ground following melt of ground ice.

Tundra – treeless terrain, with a continuous cover of vegetation.

Tundra barrens (polar semi-deserts) – areas of discontinuous vegetation in the High Arctic. The unvegetated areas may be caused by either climatic (i.e. too cold or too dry) or edaphic (i.e. low soil nutrients or toxic substrate) factors or a combination of both.

Ventifact – a stone, pebble or boulder that has been eroded or partially modified (abraded) by wind action. In cold regions, the agent of abrasion is usually either wind-blown sand particles or ice crystals. Scallops, helical scours and the formation of a surface varnish ('patina') sometimes result.