Wintertime Temperatures in the Fine-Textured Soils of the Saginaw Valley, Michigan

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Soil temperatures at four sites in Saginaw County, Michigan were monitored at 5 and 20 cm depth, on poorly-drained, loamy and clayey soils over the 1996-97 winter season. To determine the effects of land use on soil temperatures, two pairs of sites (forested and cultivated) were established. Our goal was to extend previous work that suggested that soils in Michigan freeze most frequently and for the longest duration in the Saginaw Valley area, and to present observational and quantitative data on soil freezing and fall-winter-spring soil temperatures for this region. Despite the snowy, warmer than normal winter, soils froze to depths >20 cm on cultivated sites, which were windswept and barren of snow for most of the winter, facilitating heat loss. Sites insulated by forest cover and leaf litter, as well as thin but persistent snowpacks, froze to depths of only 2 to 3 cm; temperatures at depth hovered near 1-2° C for most of the winter. Nearer the surface, soils were generally colder and had higher daily temperature variability than did soils at depth (20 cm). Forested soils were more moderated with respect to diurnal and weekly temperature change than were cultivated soils on open sites. Although both sites were located on poorly-drained soils, the wetter site developed more and larger ice lenses, and was colder throughout the fall and winter. We introduce the term 'pedothermic period' for times in which soil temperatures have consistent temporal trends and characteristics, and similar within-period variability and ranges. We identified three pedothermic periods: late fall (ends in mid December), winter (mid December to early-mid March), and spring (begins in early-mid March). Soil temperatures cool rapidly in fall, remain fairly constant under mid-winter snowpacks, but fluctuate greatly in spring after the snowpack is melted.

Keywords: Soil freezing, Michigan, pedothermic period.

This study examines soil temperatures in agricultural and forested sites in east-central southern Michigan, focusing on soil freezing. It follows a paper by Isard and Schaetzl (1998), in which the incidence of wintertime soil freezing was modelled for southern Michigan. They suggested that soils froze frequently in southeastern lower Michigan, and that this process was most frequent, intense and of longest duration in the Saginaw Valley area and the glacial lake plains (Eschman and Karrow 1985) of extreme southeastern lower Michigan (Figure 1). Our focus is on the soils of the Saginaw Valley region, which we examined during the 1996-97 winter season. We present detailed soil temperature data on the autumn 'cool down', mid-winter frozen or near-frozen conditions, and the spring "warm up". In so doing, we introduce the term 'pedothermic periods' and examine the characteristics of these periods for soils on forested sites and areas under row-crop agriculture. The aim of our study is to examine how forest clearing and row-crop agriculture have subtly altered soil thermal regimes, and hence soil hydrology, in a midlatitude, non-snowbelt location.

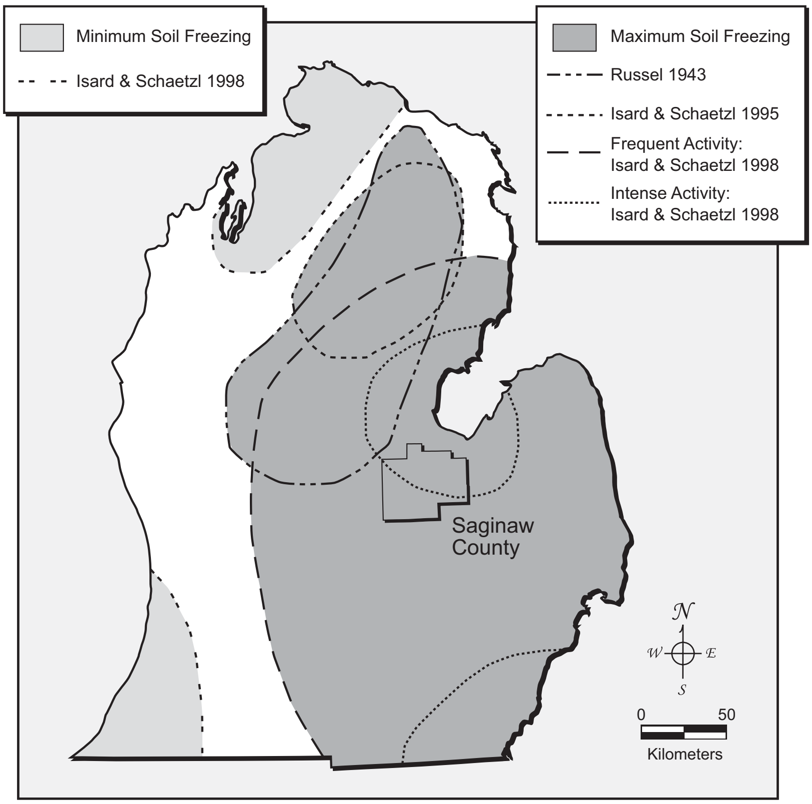

Figure 1: Areas of maximal and minimal soil freezing in southern Michigan (modified from Isard and Schaetzl, 1998).

Background

Plant growth, and biological activity, and water movement within soils are influenced by soil temperature and whether or not the soil is frozen (Post and Dreibelbis 1942, Baker 1971, Berry and Radke 1995, Sharratt et al. 1995, Carlson and Groot 1997). The rate at which heat is exchanged between the soil and the soil surface, *i.e.*, soil heat flux, ultimately controls changes in soil temperature (Oliver et al. 1987). This soil heat flux is influenced by different internal (e.g., heat capacity, soil moisture content and its effect on latent heat capacity, thermal conductivity) and external (e.g., air temperature, insulating materials at the air-soil boundary [crop and vegetative cover, snowpack, forest leaf litter]) factors (Pierce et al. 1958, Benoit et al. 1986, Johnsson and Lundin 1991, Sharratt et al. 1992, Schaetzl and Isard 1996).

Soil temperature data are important for estimating evaporation rates, mineral weathering rates, freeze-thaw processes, and frost development within soils (Stein et al. 1994). Gilichinsky et al. (1998) argued that soil, not air, temperatures should be used to gauge trends in global climate change, as they are more integrative and less volatile. Temperature also influences the movement of moisture through the soil matrix (Burt and Williams 1976). In cold regions, for example, soil frost can lower permeabilities and result in surface runoff and erosion during periods of snowmelt or rainfall (Pierce et al. 1958, Zuzel and Pikul 1987, Todhunter 2001). In addition, freeze-thaw processes can alter soil water chemistry by changing the exchangeable K^+ concentrations in soils and clay minerals (Honeycutt 1995).

Repeated freezing and thawing can affect soil structure (Post and Dreibelbis 1942) and hydraulic conductivity values (Konrad 1989), and thus the flow of solutes (Chamberlain et al. 1990). The effects of freezing and thawing on the soil environment have been addressed in studies on surface runoff (Johnsson and Lundin 1991, Stein et al. 1994), soil genesis (Schaetzl 1990, Schaetzl and Isard 1991) and soil engineering (Chamberlain and Gow 1979, Konrad 1989, Benson et al. 1995), as well as in agricultural applications (Post and Dreibelbis 1942, Zuzel and Pikul 1987). Soil freezing is especially important in areas with crops that overwinter, such as winter wheat (Larsen et al. 1988). In addition, land use practices within a watershed, such as clearing forests for agricultural purposes or other site development, can significantly alter the physical properties of soils (Chen et al. 1993). Identifying and characterizing specific physical-hydrological properties of soils, and how they vary in space and time, is therefore vital to understanding and improving upon current management practices (Aase and Siddoway 1979).

In cold regions, freezing soil conditions generally cause moisture to move upward towards the freezing front and to form ice lens structures (Radke and Berry 1998). Ice lenses are therefore more prevalent in cold wet soils than in cold dry soils. Expansion pressures, caused by water in saturated soil pores expanding as it freezes (Chamberlain et al. 1990), can heave the soil (Konrad 1989) and form additional pore space for unfrozen water to penetrate (Chamberlain and Gow 1979). As unfrozen water freezes onto the existing pore ice, still larger, crystalline ice structures can form. Over time, these ice structures may grow into areally-extensive ice lenses. In the spring, such soils may exhibit limited infiltration capacities. During freezing, large negative pore water pressures can create vertical shrinkage cracks (Chamberlain et al. 1990). Such cracks create additional pore space within the soil and increase permeabilities.

Soil frost can inhibit infiltration of snowmelt waters in early spring thereby forcing the water to pond on, or flow across, the frozen surface (Stein et al. 1994). This can potentially cause soil erosion and flooding (Zuzel and Pikul 1987, Johnsson and Lundin 1991). Groundwater recharge in the spring is partly dependent upon the amount of snowmelt water which is able to infiltrate through the vadose zone (Buttle 1989). Likewise, infiltration influences the amount of translocation of colloids and mobile organo-metallic complexes from eluvial to lower soil horizons, and therefore has a role in pedogenesis (Muir and Logan 1982, Jakobsen 1989, Schaetzl 1990).

Soils in the northern half of the United States and Canada experience wintertime freezing and thawing (Reimer and Shaykewich 1980, Zuzel and Pikul 1987, Bonan and van Cleave 1992, Schmidlin and Roethlisberger 1993). Generally, colder areas and areas with thinner, insulating snowpacks freeze more often, for longer periods of time and to greater depths (Beckel 1957, Dimo 1967, Rieger 1973, Male and Granger 1981, Ping 1987). In Michigan, the area of lowest incidence of soil freezing is associated with the lake-effect snowbelts, where deep snowpacks insulate the soil from cold, mid-winter air temperatures (Russell 1943, Schaetzl and Isard 1991, 1996, Isard and Schaetzl 1998). Sharratt et al. (1992) provided data that indicate at least 15 cm of snow are necessary to completely 'insulate' the soil and allow for 'steady state' soil temperatures. In contrast, Isard and Schaetzl (1995) concluded that snowpacks thicker than 20 cm provide little additional insulation for the soil.

Objectives

Our study of soils within Saginaw County in east-central Michigan (Figure 2) were studied during the winter of 1996-97 had three main objectives:

1. The area has comparatively low snowfall and cold winters, and has been cited as having the highest incidence of soil freezing in Michigan (Isard and Schaetzl 1998). We sought to verify or refute this (modelled) hypothesis.
2. Because most soils in the county are clayey and wet, soil freezing here should be manifested by the formation of ice lenses. The second objective was to verify that these hydrologic/pedogenic features occurred in these soils was a goal of the study.
3. The landscape is of low relief and wintertime winds are relatively unhindered on bare fields. Clearing of forests and the introduction of row-crop agriculture causes changes in snowpack thickness between forested and cultivated sites, which should result in soil temperature changes. Our third objective was to examine how soil frost incidence and soil temperatures on windswept, barren agricultural fields in winter contrasted with soils under 'stable' but thin snowpacks in adjacent woodlots (*cf* Benoit et al. 1986).

Study area

Landscapes and soils in east-central lower Michigan have primarily developed from glaciofluvial and glaciolacustrine deposits (Farrand and Bell 1982, Butterfield 1986, Monaghan et al. 1986). Most of the terrain of Saginaw County (Figure 2) has slopes of only 0 to 4% (Iaquinta 1994). Glaciolacustrine plains cover most of the county; occasionally small, early Holocene-aged sand dunes rise above the otherwise near-featureless lake plains (Arbogast et al. 1997). Most the major tributaries to the Saginaw

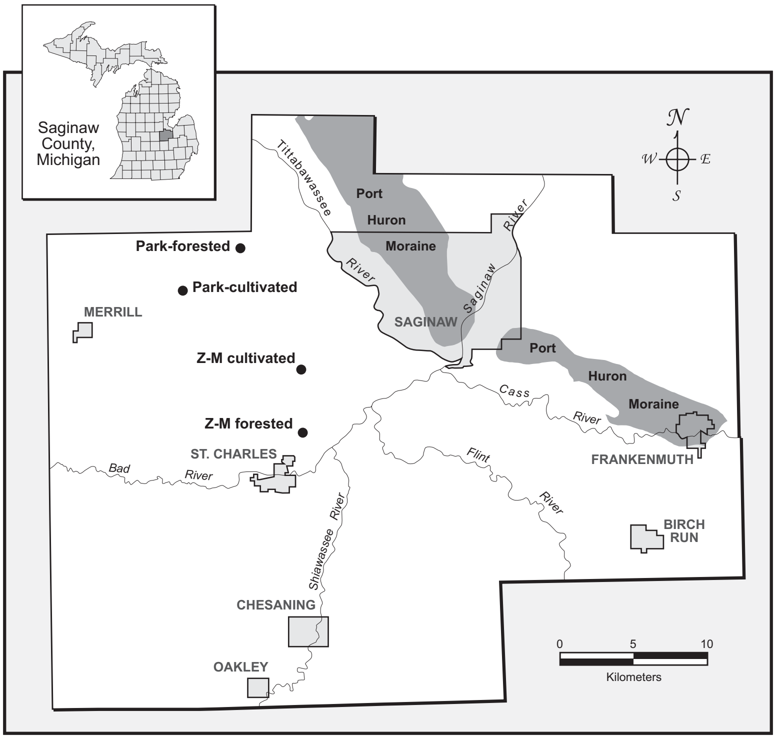

Figure 2: Study sites, cities and major landmarks/landforms, Saginaw County, Michigan.

River, which drain an area of more than 15,000 km^2, converge on a low plain in the central part of the county, immediately to the west of the Port Huron moraine (Blewett 1991; Figure 2). The outlet to this part of the watershed, through the moraine, is called the Saginaw River; it has a low gradient of approximately 1.6 cm km^{-1}. This low gradient results in very slow drainage, wet clayey soils and areas of periodic flooding, especially in areas immediately west of the Port Huron moraine.

We selected four study sites, each representative of its particular land cover type. The sites selected for this study are within the Parkhill soil series, and within a complex of Zilwaukee-Misteguay (Z-M) soils (Iaquinta 1994). Both Parkhill and Zilwaukee-Misteguay complex mapping units typify the landscapes of Saginaw County (Iaquinta 1994). Parkhill soils have loam surface horizons while soils of the Zilwaukee-Misteguay complex are dominated by silty clay textures. Typical clay contents for the Parkhill soils range from 10 to 35%, while soils of the Zilwaukee-Misteguay complex contain from 35 to 60% clay (Iaquinta 1994). Within each soil type, two locations were selected: one in a cultivated field and one in a nearby non-cultivated (forested) site (Figure 2). The cultivated sites are under no-till, row-crop (corn-soybean) agriculture.

Parkhill soils (fine-loamy, mixed, nonacid, mesic Mollic Haplaquepts; probably Orthic Gleysols in the Canadian soil classification system) are poorly drained and only moderately slowly permeable. A common profile for this series has loam Ap, Bg, and Cg horizons. An apparent water table at approximately 30 cm below the surface is common during November through May (Iaquinta 1994). The forested Parkhill site (NW$^1/_4$, NE$^1/_4$, SE$^1/_4$, Sec. 1, T12N, R2E) is approximately 10 acres in size and has various species of maple and oak (Figure 2). The cultivated site (SW$^1/_4$, SE$^1/_4$, NE$^1/_4$, Sec. 15, T12N, R2E) is about 6.5 km away.

Zilwaukee (fine, mixed (calcareous), mesic Typic Haplaquolls; probably Orthic Humic Gleysols in the Canadian soil classification system) and Misteguay (fine, mixed (calcareous), mesic Aeric Haplaquepts; probably Orthic Gleysols in the Canadian soil classification system) soils are poorly-drained and slowly permeable (Iaquinta 1994). A typical profile in this complex consists of silty clay A, Bg and Bw, and C horizons. An apparent water table at approximately 30 cm below the surface is common from October through May (Iaquinta 1994). The forested Zilwaukee-Misteguay site (SW$^1/_4$, SW$^1/_4$, NE$^1/_4$, Sec. 28, T10N, R3E) has a maple-oak forest cover. The cultivated site, located about 6.5 km away, at the Michigan State University Saginaw Valley Beet and Bean Research Farm (NE$^1/_4$, NE$^1/_4$, NE$^1/_4$, Sec. 9, T11N, R3E), is cropped to dry beans and sugar beets. Drain tiles (perforated plastic tubes) are installed at both cultivated sites to assist in soil drainage. All sites had essentially zero slope.

Methods

Soil and air temperature data were monitored at each of the four research sites from 4 December, 1996 to 31 May, 1997, using StowAway™Tidbit™-XT (Onset Computer Corporation, Pocasset, MA) automated temperature loggers with external thermocouple sensors. These instruments have an accuracy of ± 0.4 °C at +20 °C. The thin, metal thermocouple tip of each logger was inserted horizontally into the soils from the sides of small pits, at 5 and 20 cm depths and then backfilled. Temperature loggers were set to record temperatures every 72 minutes (20 readings day^{-1}). At the end of the sampling period, the temperature data were downloaded to a computer for display and analysis.

Air temperatures were also collected from each site using an Optic StowAway™ Temp (Onset Computer Corporation, Pocasset, MA) waterproof temperature logger (accuracy of ± 0.5 - 1.0 °C). The air temperature loggers were positioned near the soil temperature probes, either in a nearby tree or power line pole, at approximately 1.5 m above the surface. Each logger was housed within a one gallon plastic milk jug which had the base removed and small holes drilled into the sides to allow for free air flow, and to shade it from direct solar radiation. The loggers were set to record the temperature every 72 minutes.

Additionally, volumetric soil moisture readings were collected on-site, approximately once per month, throughout the winter using a ThetaProbe (Delta-T Devices LTD, Cambridge, England) soil moisture sensor (accuracy of ± 0.05 m^3 m^{-3}). The ThetaProbe measures the apparent dielectric constant of a soil and converts to the volumetric moisture content (θ) through a linear correlation (Roth et al. 1992, Whalley 1993). At each site, the ThetaProbe prongs were inserted vertically into three locations in the soil to instantaneously determine moisture content, at 5-10 cm and 20-25 cm depths; a mean soil moisture value was later determined.

Climatological data from the Saginaw FAA Airport and St. Charles weather stations (Figure 2) were provided by the Michigan Department of Agriculture Climatology Program at the Michigan State University Office of Climatology. Data on snowfall and snow accumulation were obtained from National Oceanic and Atmospheric Administration publications (NOAA 1996-97).

Selected properties of the soils were determined in the following summer (August, 1997), from a pit face at each site. Morphological properties, including color, structure, and consistency, within each soil profile were described according to standard methods (Soil Survey Staff 1975). Textural analysis of each horizon-based sample was determined using the pipette method (Soil Survey Laboratory Staff 1992).

Results and Discussion

Soil Physical Properties

The textures of the silt- and clay-rich, cultivated Zilwaukee-Misteguay pedon were quite similar throughout the profile; sand, silt and clay contents varied only slightly with depth, ranging from 3.5 to 5.8%, 49.2 to 51.0% and 44.1 to 46.1%, respectively. This homogeneity is attributed to the initial uniformity of the lacustrine parent materials and the lack of lessivage (clay translocation) due to longstanding high water tables. The forested Zilwaukee pedon had more notable variations in texture with depth. Clay contents ranged from a low of 39.5% in the A horizon to a high of 42.5% in the BAg horizon, indicating some translocation of clay. Low sand contents in both the cultivated (3.5 to 5.8%) and forested (4.0 to 10.2%) soils are consistent with the glaciolacustrine interpretations of these series (Iaquinta 1994).

Parkhill soils are sandier than soils of the Zilwaukee-Misteguay complex, usually with loam to clay loam textures (Iaquinta 1994). In both Parkhill soils, textures are generally sandiest near the surface and get more silty with depth, probably owing more to initial glaciolacustrine depositional processes more than to pedological processes. In the cultivated Parkhill soil, sand was the dominant particle size separate within the profile, ranging from 45.0 to 56.5%. Parkhill soils showed more evidence of clay translocation than the Zilwaukee-Misteguay soils, due to slightly lower water tables in Parkhill, under natural, undrained conditions. Clay contents, which ranged from 16.7 to 23.4%, were typically maximal in the mid-B horizon of both soils. The profile of the forested Parkhill soil was finer-textured than is the cultivated soil; its sand contents ranged from 24.1 to 40.1%, while the clay contents ranged from 34.8 to 42.4%.

Environmental Setup: Air Temperatures and Snowfall for the 1996-97 Winter

Wintertime air temperatures at nearby NWS stations indicated a cold beginning and end to the 1996-97 winter, with distinctly mild temperatures from December to March. November, 1996 was over 2 °C below normal at both Saginaw and St. Charles. Similarly, April was cooler than normal at both stations. Mid-winter (Dec-March) air temperatures, however, were above normal for both stations for all months except January at St. Charles (where it was 0.5 °C below normal).

Total snowfall amounts from November, 1996 through April, 1997, at Saginaw and St. Charles, were 195 and 138 cm, respectively. At Saginaw, every month of the 1996-97 winter season except November had more snow than the 1951-80 normals (Tomczak 2000). At St. Charles, snowfall amounts were above normal for December, January, and April (Tomczak 2000). At both locations, the largest snowfall deviation was in January, when both Saginaw and St. Charles received over twice their normal snowfall. In general, both NWS sites had measurable, more-or-less continuous, snowpacks from about 28 November to 17 March 1997 (Figure 3). Saginaw recorded its maximum snowpack of 38 cm on January 16-18, 1997, while St. Charles had a 33 cm snowpack on January 11-12, 1997.

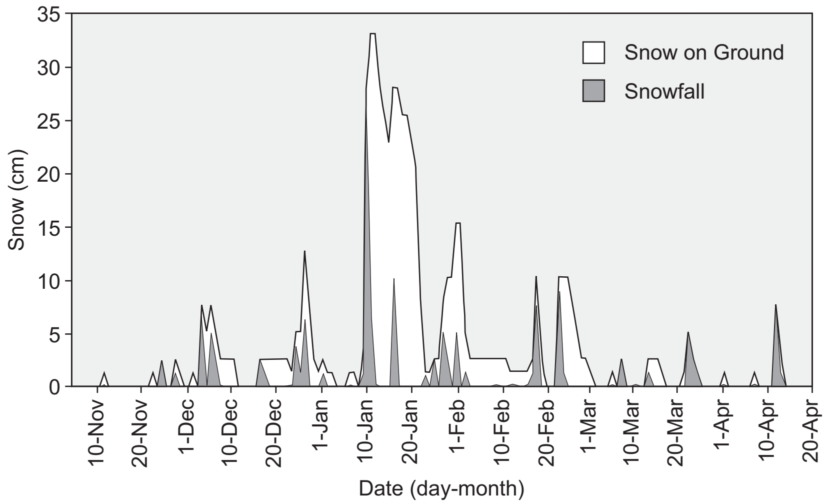

Figure 3: Snowfall and snowpack data for the Saginaw FAA airport, early November 1996 to late April 1997.

In sum, the study period should be viewed as one in which snowpacks were thicker than normal, especially during December and January, with mid-winter air temperatures that were warmer than normal. It follows that soil freezing data reported herein should therefore be viewed as minimal estimates, given the generally warmer air temperatures, and the propensity for snowpacks to insulate soils from cold, sub-freezing temperatures (Bay et al. 1952, Bocock et al. 1977, Reimer and Shaykewich 1980, Male and Granger 1981, Ping 1987, Isard and Schaetzl 1995).

Soil Temperatures

Soil temperature data were not available from the cultivated Parkhill site due to malfunctions in both temperature loggers. The presence or absence, and thickness, of frozen soil at the cultivated Parkhill soil was therefore estimated visually.

At the start of the study period (4 December), soils at all sites and depths had cooled to approximately 4 °C (Figure 4). Continued cooling resulted in a temperature plateau (more-or-less) that persisted throughout the winter months. This 'plateau' was between 0 and 2 °C for most sites and depths, with occasional freezing 'spikes' in February at the cultivated sites (Figure 4). By mid-March, loss of snowpack and warming air temperatures led to gradual warming of the soils, although the highly fluctuating nature of the spring soil temperatures reflected variability in air temperatures during this transition season.

Near-surface soils obtained overall colder mid-winter temperatures, and were more variable in time, than were soils at 20 cm depth (Figure 4). Warm 'spikes', such as the mid-January spike (Figure 4), were present but more subdued at 20 cm than at 5 cm. During winter (mid-December to early March), soil at 5 cm was an average of 0.7 °C colder than the soil at 20 cm at both the cultivated and forested Zilwaukee-Misteguay sites (-0.8 °C at the forested Parkhill site). Within the cultivated Zilwaukee-Misteguay complex, soils reached a low temperature of -2.7 °C at 5 cm on February 20 and -0.4 °C at 20 cm from February 20-22. This temporal variability is not unexpected (e.g. Goetz and Müller 1969). For example, on March 30, 1997, the soil temperatures at the Zilwaukee-Misteguay cultivated site fluctuated over a 24 hour period from 2.7 to 8.8 °C at 5 cm, but only from 2.6 to 4.3 °C at 20 cm. Deep soil horizons are more effectively insulated by overlying soil material and are less affected by fluctuating air temperatures than are near-surface horizons (Smith et al. 1964, Ping 1987, Braley and Zarling 1991). Cold air temperatures contributed to colder temperatures at 5 cm,

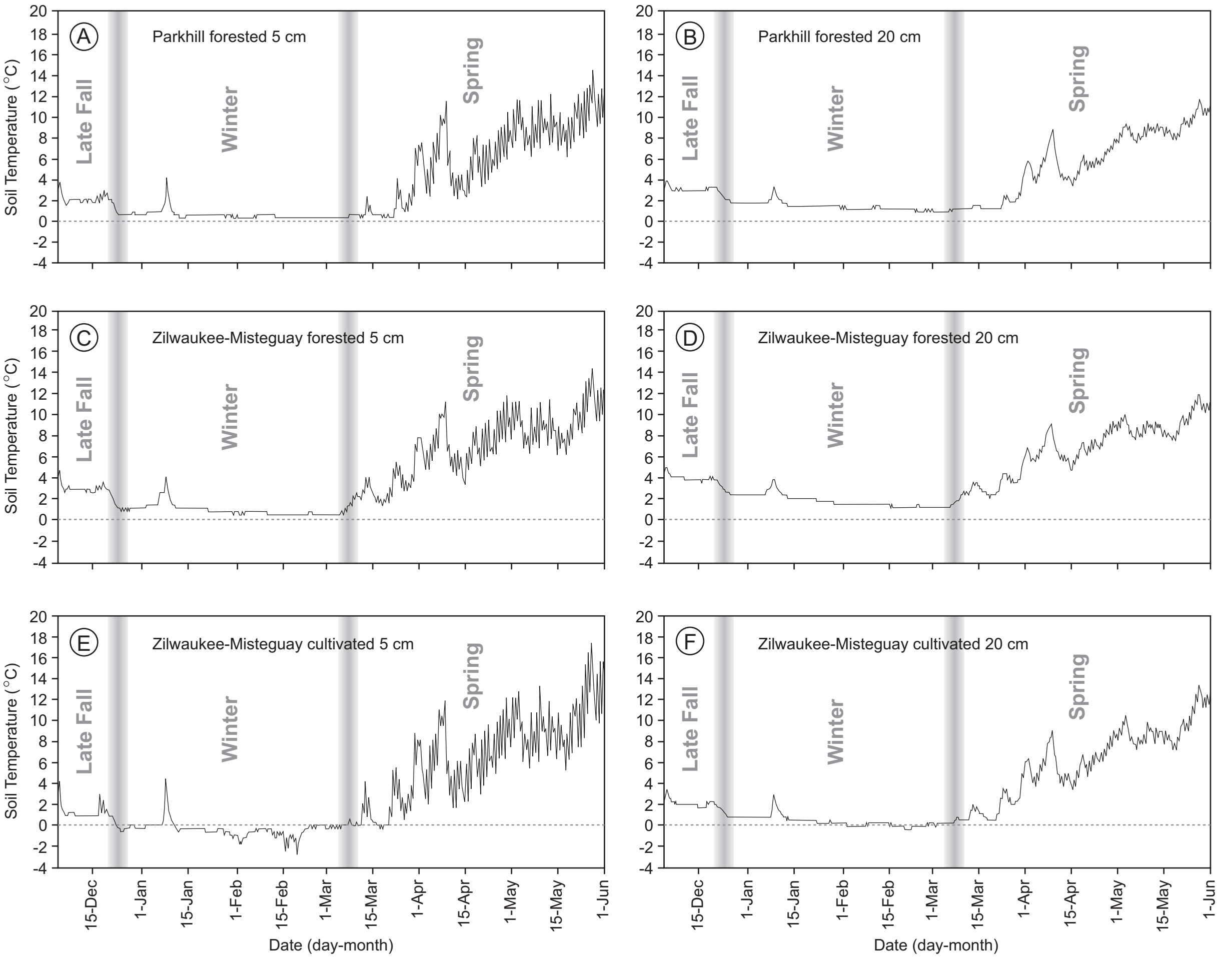

Figure 4: Soil temperatues at the Zilwaukee-Misteguay (forested and cultivated) and Parkhill (forested only) sites at 5 and 20 cm. Approximate beginning and ending dates of the three soil pedothermic periods are shown.

while soils at 20 cm are buffered by overlying soils and respond more slowly to the colder air temperatures (Isard and Schaetzl 1995).

Forested vs Cultivated Soil Temperatures

Soils at both forested sites did not freeze at 5 cm, although frost was observed down to 2-3 cm (Figure 4). Soils at the cultivated Zilwaukee-Misteguay site froze to as deep as 20 cm and remained frozen for over a week in late February (Figure 4). Although soil temperature data were unavailable for the cultivated Parkhill site, frozen soil and frost lenses were observed *in situ*. Colder temperatures and frost at the cultivated sites was not unexpected, since these sites had lain windswept and barren for much of the winter (see also Pikul Jr. et al. (1986) and Benoit et al. (1986)). Soils studied in open fields in the Saginaw Valley freeze and stay frozen for significant periods of time, and to 20+ cm depth, even in winters when the incidence of soil freezing may be lower than normal, as was the case in the winter of 1996-97.

Wooded cover and soil litter layers (O horizons) at the forested sites insulated the soil from cold air temperatures (Benoit et al. 1986, Pikul et al. 1986, Stein et al. 1994, Kite 1998), allowing for warmer and less variable temperatures than in nearby cultivated sites (Figure 4). Snow cover can also limit the extent of heat loss from the soil to the atmosphere, and therefore limit the extent to which the soil temperatures decrease in midwinter (Benoit et al. 1986, Johnsson and Lundin 1991). Thus, deforestation and cropping of this landscape appeared to have jointly led to lower wintertime soil temperatures, an increased

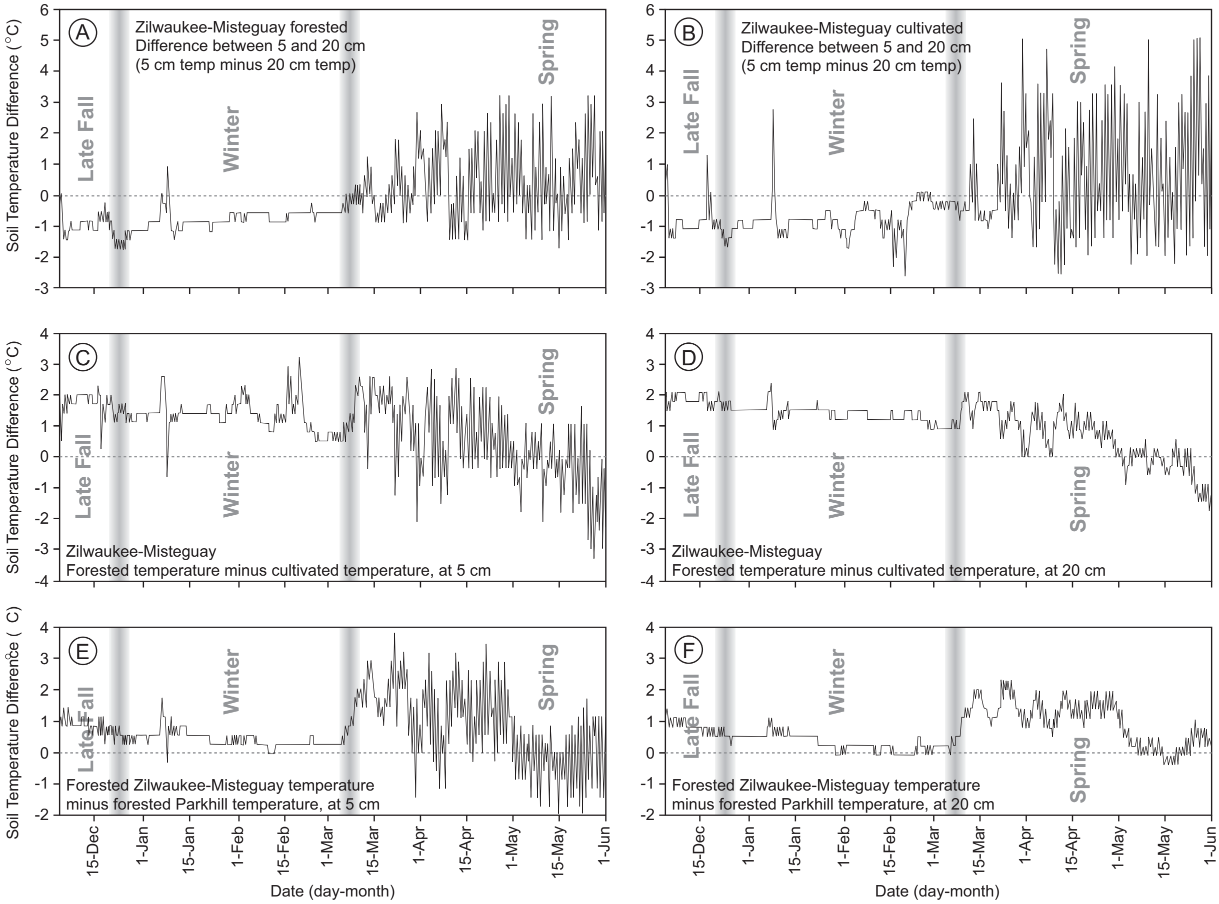

Figure 5: Soil temperature differential between the Zilwaukee-Misteguay (forested and cultivated) and Parkhill (forested only) sites at 5 and 20 cm. Approximate beginning and ending dates of the three soil pedothermic periods are shown.

incidence of freezing, and freezing to greater depths than in the "naturally" forested landscape.

We further examined temperature differences between the forested and cultivated sites of the Zilwaukee-Misteguay complex at 5 and 20 cm, to ascertain the specific effects of deforestation and cultivation of row crops (which are absent in the winter, unlike forage crops) on soil temperatures. Throughout the late fall and winter, soil temperatures at both depths were almost always higher at the forested site (Figure 5). During midwinter, mean soil temperatures at the forested site were approximately 1.3 °C (5 cm) and 1.4 °C (20 cm) higher than at the cultivated site (Figure 5). This is interesting, given that during the study period, mean air temperatures at the cultivated sites were consistently higher than at their forested pair (Tomczak 2000). In a similar study comparing a clear-cut region with a closed canopy forest, it was concluded that higher average daytime air temperatures within the clear-cut region were related to greater irradiance (Carlson and Groot 1997). With the onset of spring, soil temperatures at both depths began to fluctuate (Figures 4, 5).

Soil temperature variability (daily, seasonally, etc.) was also lower at the forested sites (Figure 4), illustrating the moderating or buffering effect of forest (*cf* Qashu and Zinke 1964, Oliver et al. 1987, Balisky and Burton 1993). In general, the open cultivated sites also experienced both higher maxima and lower minima air temperatures than did the forested sites. For example, in May 1997 the maximum-minimum air temperature differences at the cultivated and forested Zilwaukee-Misteguay sites were 41.6 and 28.3 °C, respectively. Carlson and Groot (1997) reported similar data (17.1 and 10.1 °C) in a clear-cut and nearby forested setting in summertime. In addition, Chen et al. (1993) reported maximum-minimum air temperature differences of 14.8 and 10.1 °C in a clear-cut area and an adjacent forest.

Comparative Soil Temperatures

Air temperature comparisons between the two forested sites were made to determine climatic differences between unlike soils with the same land use (Figure 4). In general, the forested Zilwaukee-Misteguay site was warmer than the forested Parkhill site for all of the study period with the exception of late spring. In mid-winter, temperatures at 20 cm were comparable for short periods of time, while at the same time the forested Parkhill site was about 0.5 °C warmer at 5 cm depth. The period of greatest temperature difference occurred in March, during snowmelt, when the wetter Z-M site was occasionally ponded. The preponderance for wet soils to be colder than dry soils might explain the temperature difference between these two sites, and the two series.

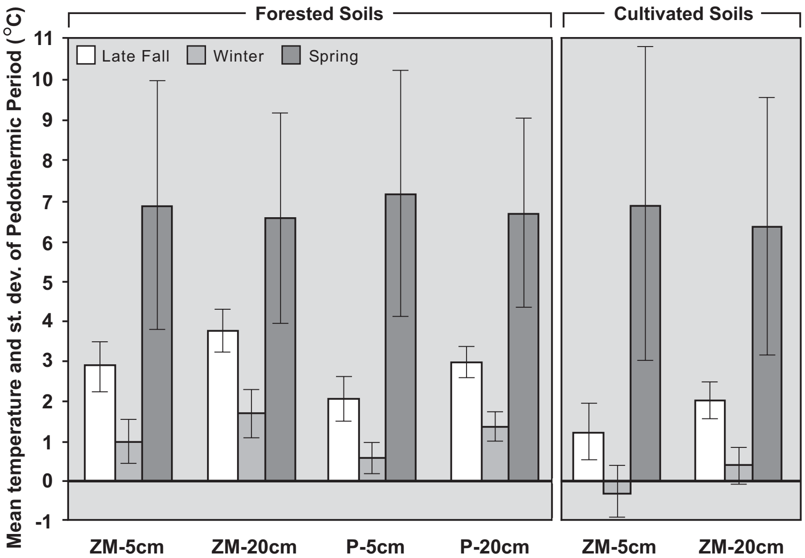

Figure 6: Mean soil temperatures and standard deviations at six sites for the three pedothermic periods in the Saginaw Valley area.

Soil Pedothermic Periods

Periods of time with internally-consistent temperature patterns or trends, with similar within-period variability and ranges, were qualitatively estimated from the soil temperature plots (Figure 4) and referred to as "pedothermic periods". Three distinct pedothermic periods were determined for the Saginaw area: (1) Late Fall (ends in mid December), Winter (mid December to early-mid March), and Spring (begins in early-mid March). Other regions may have different pedothermic periods, or the same periods with different starting and ending dates.

Steadily decreasing soil temperatures, with low daily variability, were characteristic of Late Fall (Figures 4, 6). Since data collection did not begin until December 4, 1996, it is not known when the Late Fall period began. During Late Fall, variability in soil temperatures was low, with standard deviations ranging from 0.39 °C (20 cm forested Parkhill) to 0.70 °C (5 cm cultivated Zilwaukee- Misteguay) (Figure 6). Low soil temperature variability is attributable to the insulating effects of the abnormally-thick snowpack (Benoit et al. 1986, Sharratt et al. 1992, Schaetzl and Isard 1996), which was the result of above normal snowfall during December, 1996 (Figure 3). In other Fall periods, variability could and probably would be higher.

The Winter pedothermic period (mid December to early-mid March) began around December 23-25, and coincided with the first prolonged, sub-freezing event on December 22-24. During these three days, minimum daily air temperatures ranged from –14 to –16 °C, decreasing soil temperatures by as much as 1.8 °C (5 cm forested Z-M) before reaching a steady state. During the Winter pedothermic period, mean soil temperatures stayed consistently near or slightly above freezing, with low variability. Standard deviations of Winter soil temperatures were low, ranging from 0.37 °C (20 cm forested Parkhill) to 0.68 °C (5 cm cultivated Z-M), again attributable to snow cover (Figure 6). Mean soil temperatures ranged from –0.3 (5 cm cultivated Z-M) to 1.7 °C (20 cm forested Z-M). The change of state from liquid water to ice releases latent heat, which forces the actual temperature of a cooling soil to hover near freezing for some period of time. This process is illustrated in Figure 4; here, the forested Parkhill soil cools to near the freezing point at 5 cm. Above, at 2-3 cm depth, the soil was frozen. Thus, the release of latent heat, which occurred as soil water froze, "warmed" the soil at 5 cm, holding it near 0^{o}C. At shallow (5 cm) depths, the cultivated soils remained frozen during the Winter period, indicative of the effects of cold air temperatures on open, wind-blown fields (Benoit et al. 1986, Pikul et al. 1986). In these soils, the freezing front had penetrated below 5 cm; once frozen, the soil could cool to sub-zero temperatures, as it lacked further additions of latent heat.

The exact end date of the Winter pedothermic period depends upon the site and undoubtedly varies from year to year, but for this winter ranged from early to mid March. The end of the Winter pedothermic period is indicated when soil temperatures change from near steady to generally increasing values, and increasing degrees of variability. The end of the Winter period coincided, in these cases and (we suspect) in many others, with the loss of snowpack from the sites. Pierce et al. (1958) reported that cover in the forests can slow the warming of soils in the spring, thereby enabling frost (*i.e.*, Winter pedothermic) conditions to persist.

Although snowpack thickness varied throughout the 1996-97 winter, the period with the greatest snowpack thickness was from early December to late February (Figure 3). The lag between the end of the persistent snowpack (March 2) and the end of the Winter period (mid-March) is not unusual, and reflects stored latent heat within the wet soils. For example, Stein et al. (1994) reported that soil temperatures (at 6 cm) in Quebec took 5-25 days to warm to above 0 °C after the snowpack had melted.

Although other snowfall events occurred after March 2, the snowpack never reformed for long periods of time (Figure 3). During the first week in March, daily air temperatures reached 12 °C at the cultivated Zilwaukee-Misteguay site, signaling the end to any Winter-like soil temperatures; the transition from the Winter to the Spring pedothermic period had started.

The third pedothermic period we identified, Spring, began in mid March. Mean soil temperatures had high rates of change (warming) and extremely high variability during this period, which lagged behind the warming atmosphere. Mean soil temperatures increased significantly over the mean Winter soil temperatures, by as much as 7.2 °C at 5 cm of the cultivated Zilwaukee-Misteguay site. In a similar study, Sharratt (1993) reported that mean minimum soil temperatures at 5 and 20 cm increased 6.9 and 7.9 °C, respectively, between April and May, after the loss of the snowpack. After late March, or the start of the Spring pedothermic period, soil temperatures at 5 cm became warmer than at 20 cm during the daytime hours for the forested Parkhill site. Later in Spring, as warming of the soils continued from above, soils at 5 cm were often warmer than the soils at depth, although diurnal variability was great (Figure 4).

Standard deviations were very high during the Spring pedothermic period, ranging from 2.4 °C at 20 cm of the forested Parkhill site to 3.9 °C at 5 cm of the cultivated Zilwaukee-Misteguay site. This trend is attributable to the absence of a snowpack and cover crop, exposing the soils to fluctuating daily air temperatures (Stein et al. 1994). Because our data collection had ended on May 31, we do not report on any ensuing pedothermic periods, or what characteristics trigger the end of the Spring pedothermic period.

Soil Frost and Ice

Frost and ice lenses were observed within the soil at both cultivated sites in February and March. In February, characteristic concrete frost had formed to 15-20 cm at the cultivated sites (Zuzel and Pikul 1987). By March, this frost had thawed in the upper 12-15 cm, but was still present below 15 cm at both cultivated soils. At both forested sites, frost was only observed near the surface and, in these cases, was more porous and granular.

The open, wind-swept fields of both cultivated sites had minimal snowpacks, exposing the soils to cold air temperatures and enabling concrete frost to penetrate deep within the Ap horizon. Massive concrete frost can eventually become nearly impervious. Occasional ponding and runoff can then occur (Mostaghimi et al. 1988, Stein et al. 1994). Conversely, the thin, granular frost present at the forested sites did not appear to impede snowmelt infiltration (Zuzel and Pikul 1987, Schaetzl 1990).

Ice lens structures generally develop in soils that have fine textures and high moisture contents (Chamberlain and Gow 1979; Kay et al. 1985), such as the soils studied here. Although identified in both soil types, ice lenses were more prominent at the Zilwaukee-Misteguay site, probably because the Parkhill sites are drier. The Zilwaukee-Misteguay soil was wetter in 10 of the 12 paired soil moisture readings taken with the ThetaProbe (e.g., 5 cm forested Parkhill vs 5 cm forested Z-M). On average, the Z-M site soils contained 118% of the soil water, by volume, of the comparable Parkhill sites (Tomczak 2000). The macromorphology of the ice lenses observed with the Zilwaukee-

Misteguay soil was similar to those found elsewhere on clay loam (Kay et al., 1985) and other clay soils (Chamberlain and Gow, 1979), ranging from 2-8 mm in thickness and up to 10 cm in length. Parkhill lenses ranged from 4-8 mm in thickness and up to 4 cm in length. Ice lenses in the Parkhill soil generally developed as singular structures throughout the soil matrix, whereas lenses in the Zilwaukee-Misteguay soil were linked together and interconnected polygonally.

In large part due to the frozen soil conditions at the cultivated sites, snowmelt and spring rains led to saturation of the surface soils by March; ponding in swales of the forest floor was also evident on both forested soils. Depending on the yearly extent of snowmelt and spring rain events, such ponding could lead to localized flooding. Conversely, deep frost and ice lens formation can break up and expand the soil matrix, creating additional macroscopic pore space and increasing permeability (Kay et al. 1985, Chamberlain et al. 1990). In the spring, reconsolidation of the soils may only remove some of the newly formed pore space. Macroscopic cracks remaining after the melt period can function as channels of low resistance to water flow.

Conclusions

Isard and Schaetzl's (1998) study had predicted intense soil freezing in forested, upland soils of the Saginaw Valley. The winter of 1996-97 did not exhibit such conditions at forested sites, (although deep freezing occurred at cultivated sites), due to snowpack and air temperature conditions that interacted to reduce the incidence of freezing, and because the wet soils studied here are less likely to freeze than are the upland soils modelled by Isard and Schaetzl. Nonetheless, our study showed that soils did freeze in this region, but less so on wetter sites. The effects of deforestation and subsequent row crop cultivation on the 'natural' soil temperatures are especially noteworthy, since cultivated soils in this region froze to nearly ten times the depth as than did forested soils.

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