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Southern Michigan's Snow Belt
Modelled Impacts on Soil Temperatures and Freezing

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ABSTRACT

A physically-based computer model is developed for estimating soil temperatures at 0.05, 0.1, 0.2, and 0.5 m depths in sandy, forested soils, using daily minimum and maximum air temperatures and precipitation as input data. The algorithm is evaluated by comparing its output to 490 soil temperature observations for 1990-93 at stations in southern Michigan, USA. The mean errors of the soil temperature estimates are less than 0.1°C at the 0.5 m depth, thus comparing favorably to models requiring more detailed input. The model is applied to 14 stations across lower Michigan, using 1951-91 climatic data, in order to examine regional trends in soil temperature and freezing and to compare these trends to patterns of snow thickness and air temperature. Simulations reveal that soils seldom freeze within the deep snow area (lake-effect snow belt) of southern Michigan and along the coast of Lake Michigan. Soil freezing is more common at non-snowbelt, interior locations.

INTRODUCTION

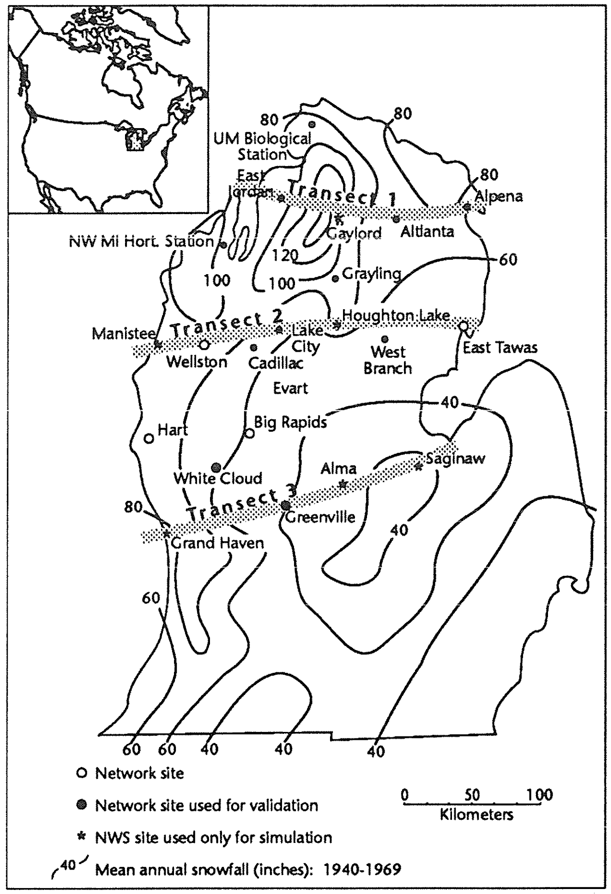
In this paper we present soil temperature calculations for southern Michigan, USA, generated by a computer algorithm that uses as input daily minimum and maximum air temperature and precipitation collected by the US National Weather Service (NWS). This region was chosen because it experiences sub-freezing temperatures and has widely variable differences in snow accumulations, both spatially and temporally. The algorithm computes soil temperatures at multiple depths, as well as snowcover thicknesses. The calculations are verified from actual soil temperature measurements. The model is then used to simulate soil temperatures over 41 years (1951-1991) at 14 NWS sites across the region.

MODEL CONSTRUCTION

A computer algorithm was developed to calculate the vertical profile of temperature for coarse-textured, well-drained, forested soils. The model also has a water balance component that calculates infiltration of water through the litter layer and into the mineral soil (see Schaetzl and Isard [1990] for more details). The model combines a Thornthwaite-based water budget model (Thorn-thwaite and Mather 1955) and the SCS's runoff and snowmelt models (USDA-SCS 1971) with a one-dimensional heat conduction equation (Carslaw and Jaeger 1959). The model is driven by daily maximum and minimum air temperatures and precipitation data obtained by the NWS. Precipitation that falls when the mean daily temperature (MDT) is ≥0°C is assumed to be rain; other precipitation is snow. The algorithm uses a 10:1 conversion for liquid to solid precipitation. The water balance part of the model is discussed in detail elsewhere (Schaetzl and Isard 1991). Snowmelt is calculated as a function of mean daily air temperature (MDT, Garstka 1964) and is made available for storage in the litter layer.

The temperature profile within the mineral soil is calculated at 30 min intervals using a finite difference formulation with nine soil layers that increase

*Figure 1. Map of southern Michigan showing locations of US National Weather Service stations used in this study. The soil temperature network (monitoring) sites are indicated with circles; hollow circles represent those used for model verification purposes. NWS sites that were used in the analysis of 1951-91 soil temperature trends are located with a *. They form three east-west transects across southern Michigan. Transect 1 is from East Jordan to Alpena. Transect 2 is from Manistee to East Tawas, exclusive of Wellston. Transect 3 is from Grand Haven to Saginaw. Isolines of mean annual snowfall are overlain across the grid of stations.*



in thickness with depth (0.05, 0.1, 0.3, 0.5, 1.0, 1.5, 3.5, and 8.0 m thick, respectively), one litter layer (0.05 m thick), and up to five snowpack layers, all 0.1 m thick. The soil temperature at 15 m is held constant at 2°C above the mean annual air temperature, as suggested by Smith et al. (1964). Daytime temperatures at the surface for the 30 min intervals are expressed as a truncated harmonic function of time around the MDT; an exponential function is used to calculate nighttime temperatures (Parton and Logan 1981). Thermal properties for the soil, litter, and snowpack are taken from van Wijk and de Vries (1963). Thermal properties of the soil layers vary as a function of the computed water content. The only site-specific input parameters needed to operationalize the algorithm are latitude, forest type (coniferous, deciduous, or mixed), water retention capacity of the litter layer, SCS hydrologic soil group classification, and SCS forest condition classification.

MODEL VERIFICATION AND ACCURACY

A soil temperature measurement network was established in 1990 across southern Michigan; 14 sites are currently operational (Fig. 1). Copper-

Table 1. Error Statistics for soil temperature model.

Depth (m)	r^2	*RMSE (°C)*	*d*	*MBE(°C)*	$(RMSE_s/RMSE)^2$
Annual Series (n=490					
0.05	0.93	2.3	0.97	-0.4	0.20
0.10	0.94	2.1	0.98	-0.4	0.20
0.20	0.94	1.8	0.98	-0.2	0.17
0.50	0.93	1.3	0.98	-0.1	0.02
Warm Season (May–October) Series (n=261)					
0.05	0.83	1.9	0.94	+0.6	0.12
0.10	0.83	1.9	0.94	+0.6	0.11
0.20	0.84	1.7	0.94	+0.6	0.10
0.50	0.82	1.5	0.95	-0.1	0.05
Cool Season (November–April) Series (n=229)					
0.05	0.47	2.5	0.73	-1.6	0.52
0.10	0.53	2.3	0.75	-1.5	0.55
0.20	0.63	1.8	0.83	-1.1	0.47
0.50	0.80	1.1	0.95	0.0	0.03

constantan thermocouples were installed at each location, 0.05, 0.1, 0.2, and 0.5 m below the mineral soil surface. Soil temperature at each depth, time of observation, and snowpack thickness (mean of three readings) are recorded bimonthly at each site. All sites are located within 16.3 km of a NWS station, are forested, and on upland, nearly level landscape positions with well-drained, sandy soils. Data from sites with more than 40 observations (10 of the 14 sites) were used to evaluate the accuracy of the soil temperature computations.

Table 1 provides error statistics for the soil temperature model—stratified by depth and into annual, warm, and cold season categories. The coefficient of determination (r^2) for the annual series (all observations) is ≥ 0.93 for all depths. Predictions of soil temperature are more accurate for the warm season (May-October) than for the cold season (when snowpacks are usually present). r^2 values for all depths in the warm season are ≥ 0.82, whereas cool season values range between 0.47 and 0.80. The r^2 values vary little with depth for the warm season. In contrast, the strength of the relationship between predicted and observed soil temperatures increases dramatically with depth for the cool season. Variations in model accuracy between seasons and with depth during the cool season are primarily related to the computation of snowpack thickness. Likely causes of the errors associated with this component of the algorithm are discussed below.

Inspection of the root mean square error statistics (RMSE) reveals that the expected errors of computed soil temperature values decrease with depth, from 2.3°C at 0.05 m to 1.3°C at 0.5 m for the annual series (Table 1). This corresponds to a decrease in diurnal variations in soil temperature with depth. The index of agreement (d) for both the annual and warm season series suggest that approximately 95% of the "potential" for error in predicting instantaneous soil temperatures at each of the four depths is "explained" by the simulated soil temperatures. Model predictions for the cool season are generally less accurate, except at the 0.5 m depth where the RMSE is only 1.1°C. The model generally underestimates soil temperatures at the 0.05, 0.1, and 0.2 m depths for the annual series (mean bias errors, MBE, are between -0.2 and -0.4°C). For the warm season, soil temperatures at these depths are overestimated by 0.6°C; temperatures are underestimated by 1.1 to 1.6°C for the cool season. The MBEs at 0.5 m are very small (± 0.1 °C) regardless of the season. In general, less than 20% of the expected error in the annual and warm season predictions (i.e., $(RMSE_s/RMSE)^2 \leq 0.2$) can be attributed to systematic causes, probably contained within the model. The systematic error is ≤ 5% of the total error for the 0.5 m level. The large systematic errors (≥ 50% of the total error) for the upper layers during the cool season support the contention that the snowpack thickness component is an important source of model error.

Calculated vs observed soil temperatures for the 0.05 and 0.5 m depths are plotted in Figure 2. The summer "warm bias" (overestimation of warm temperatures) and the winter "cold bias" (underestimation at low temperatures) of the model are

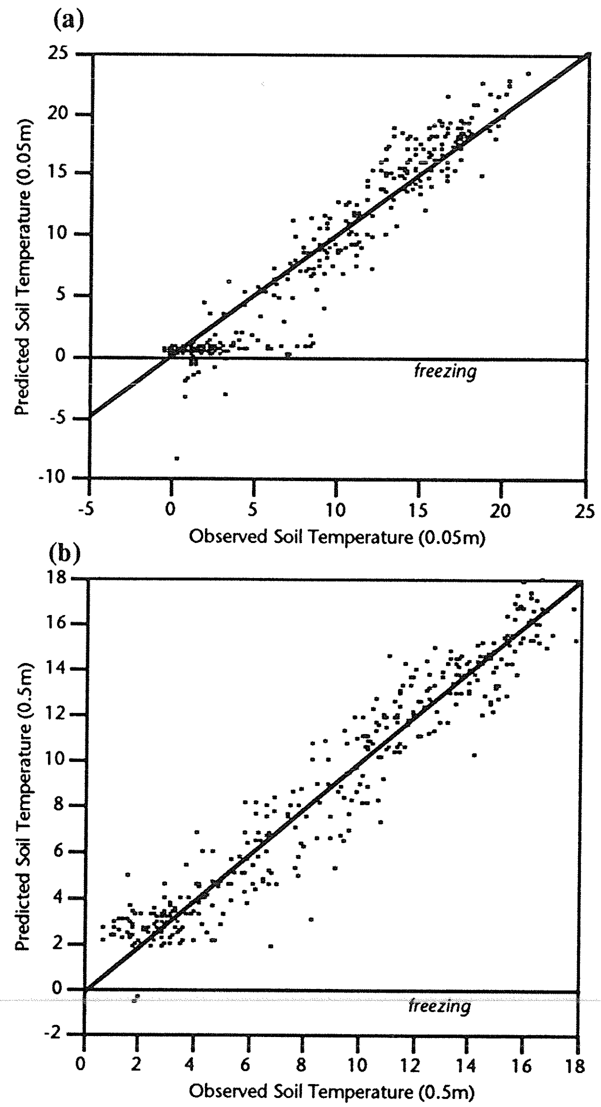


Figure 2. Relationship between observed and predicted soil temperatures at the 0.05 m (a) and 0.5 m (b) depths for all measurement sites. The 1:1 correspondence is shown as a diagonal line—it is not the least squares regression line.

Table 2. Soil temperature model error statistics for individual stations. Annual series averaged for all depths.

Station Name	*n*	r^2	*RMSE (°C)*	*d*	*MBE °C)*	$(RSE_s/RMSE)^2$
Pellston	232	0.96	1.4	0.99	+0.3	0.15
Atlanta	216	0.93	2.6	0.96	-1.4	0.43
East Jordan	164	0.94	1.8	0.98	+0.4	0.30
Lake City	196	0.90	2.0	0.97	-0.1	0.01
West Branch	200	0.94	2.1	0.97	-0.1	0.36
Greenville	200	0.93	1.9	0.98	0.0	0.15
NW Michigan Hortic. Station	236	0.94	2.0	0.97	-0.2	0.34
Alpena	200	0.96	1.5	0.99	-0.6	0.22
White Cloud	188	0.96	1.5	0.99	-0.1	0.12

apparent at 0.05 m, but not at 0.5 m (Fig. 2). "Warm bias" likely occurs because differences in air temperatures, especially on sunny days, between NWS instrument shelters in open areas and neighboring shaded (and cooler) forest sites are greater during summer than in the other seasons.

Snowpacks under forest cover accumulate and melt more slowly than do those in the open (Baldwin 1957). The weak association between the calculated snowpack depths and those observed at the forested sites (not shown here) illustrates that the algorithm often appears to melt the snowpack too rapidly in general, as well as on days when the MDT is slightly greater than 0°C at the NWS shelter but likely below freezing in the forest. As a result, the largest differences between simulated and observed soil temperatures occurred for cold winter days that follow periods with temperatures ≥0°C. Because the calculation of melt is too great for many warm winter days, the simulated snowpack is too thin to provide the modelled soil with enough "insulation" from cold air temperatures on following days, predicting upper soil layers that are much colder than observed. The importance of this source of error in the soil temperature calculations decreases with increasing depth because the soil between the measurement depth and the surface provides additional "insulation" from cold air temperatures. This shortcoming of the model is in large part due to the simplicity of the linear function that is used to calculate snowmelt from mean daily air temperature. Certainly, many other atmospheric factors have been shown to have important impacts on snowmelt including radiation, wind, and humidity. Other limitations of the snowpack component of the model are that it uses a constant to convert liquid precipitation to snow and does not account for changes in snow density and thermal properties as the snowpack ages or "ripens".

The model was also evaluated to determine the sensitivity of the soil temperature computations to changes in the water budget input parameters (stemflow and throughfall coefficients, evaporation formulation, water retention capacity of the litter, hydrologic soil group classification, SCS forest condition). As expected, the water budget component of the model was altered by changes in the above input parameters (see Schaetzl and Isard [1990] for a comparable sensitivity analysis on an earlier version of the model), but the changes had little effect on the soil temperature calculations. Baker (1971) also found that the influence of soil moisture on soil temperatures was insignificant when compared to the effect of snow cover.

Table 2 shows the soil temperature model errors associated with selected individual stations, aggregated for all depths and seasons. The errors for the individual stations, with the exception of Atlanta (Table 2), are very similar to those for the combined data set (Table 1), illustrating that the model is not site-specific, and can be applied at many forested, sandy sites. In general, stations for which the systematic error is a large proportion of the total error had more snowcover during the measurement period than the other sites. The RMSE for the stations compare favorably with the expected errors of 1° to 3°C associated with a soil temperature model recently developed by Vose and Swank (1991). In that study, mean daily soil temperatures for a snow-free, forested site in North Carolina were simulated using hourly air temperature inputs.

REGIONAL SOIL TEMPERATURE SIMULATIONS

The effect of snowpack thickness on air-soil temperature relationships and freeze-thaw frequency was analyzed using data from 14 NWS stations located in three east-west transects across the lake-effect snowbelt in southern Michigan (Fig. 1). Southern Michigan's lake-effect snow belt occurs in a north-south tract approximately 40 km inland from Lake Michigan (Eichenlaub 1970). Snowfall in the

Table 3. Correlation coefficients between monthly mean air and soil temperatures (n=574 obs.).

Depth (m)	*Jan*	*Feb*	*Mar*	*Apr*	*May*	*June*	*July*	*Aug*	*Sep*	*Oct*	*Nov*	*Dec*
0.05	0.44	0.28	0.52	0.89	0.97	0.98	0.99	0.99	0.98	0.98	0.91	0.57
0.10	0.43	0.28	0.50	0.87	0.95	0.96	0.97	0.97	0.96	0.97	0.90	0.59
0.20	0.41	0.27	0.43	0.84	0.88	0.89	0.93	0.93	0.91	0.93	0.86	0.61
0.50	0.35	0.29	0.22	0.75	0.66	0.70	0.81	0.82	0.81	0.80	0.76	0.51

Table 4. Mean monthly air temperature and computed soil temperatures (0.05 m depth) for stations forming a transect across the snowbelt of southern Michigan between 1951 and 1991.

	Jan	*Feb*	*Mar*	*Apr*	*May*	*June*	*July*	*Aug*	*Sep*	*Oct*	*Nov*	*Dec*	*Annual*
Air Temperature (°C)													
East Jordan	-6.3	-6.4	-1.5	6.0	12.4	17.2	19.8	18.9	15.0	9.5	2.9	-3.4	6.1
Gaylord	-8.0	-7.3	-2.4	5.7	12.4	17.2	19.7	18.6	14.3	8.6	1.4	-5.2	6.3
Atlanta	-8.0	-7.3	-2.2	5.5	12.1	16.8	19.5	18.4	14.1	8.4	1.7	-5.0	6.3
Alpena	-7.6	-7.1	-2.4	5.1	11.3	16.5	19.4	18.3	14.0	8.4	1.9	-4.5	6.1
Soil Temperature (°C) at 0.05 m													
East Jordan	0.1	0.2	0.3	2.3	8.1	11.8	13.8	13.3	10.8	7.1	2.7	0.2	5.3
Gaylord	0.1	0.0	0.2	1.9	7.9	11.8	13.8	13.3	10.8	6.6	2.1	0.1	5.7
Atlanta	-0.5	-0.1	0.3	2.4	7.9	11.7	13.6	13.0	10.4	6.6	2.2	-0.3	5.6
Alpena	-0.5	-0.1	0.2	1.8	7.2	11.3	13.4	13.0	10.3	6.4	2.3	-0.2	5.4
Air–Soil Temperature Difference (°C)													
East Jordan	-6.5	-6.6	-1.8	3.7	4.4	5.4	6.0	5.5	4.0	2.4	0.2	-3.6	
Gaylord	-8.1	-7.3	-2.6	3.8	4.5	5.3	5.9	5.4	3.5	2.0	-0.7	-5.2	
Atlanta	-7.5	-7.2	-2.5	3.2	4.2	5.2	5.9	5.3	3.7	1.8	-0.6	-4.7	
Alpena	-7.1	-6.9	-2.7	3.3	4.2	5.2	6.0	5.4	3.7	2.0	-0.4	-4.3	

month or season of record. In order to characterize regional patterns of soil temperature and freezing across southern Michigan, many of these raw outputs were averaged over the 41 year record for each station to examine their impact on soil formation.

AIR-SOIL TEMPERATURE RELATIONSHIPS

Correlation coefficients for relationships between mean monthly air and soil temperatures (Table 3) show strong positive correlations ($P < 0.001$) between air and soil temperatures for all months at all depths. The relationships are especially strong for the upper soil layers during snow-free months, when the lack of snow cover allows these layers to respond quickly to air temperature fluctuations. Lower correlations between soil temperatures at 0.2 m and 0.5 m and air temperatures in all months point to the insulating effects of the soil above. Because temperatures in deeper soil layers respond more slowly to air temperature fluctuations (a "lag effect"), they are often correlated with air temperatures of the previous month. For example, the correlation between air and soil temperatures for October is 0.80 at 0.5 m. When the air temperatures for October **and** September are both used in the regression to predict soil temperatures at 0.5 m, the multiple correlation increases to 0.93. The relationship, as indicated by the r values, increases very little or not at all when the previous month's air temperatures are added to the regression equation for winter and summer months, because changes in air and soil temperature from one month to the next during summer and winter are not as large as in the "transition" seasons. In addition, the influence of the previous month's air temperature on soil temperature during winter is overshadowed by the insulating effects of the snowpack.

Air-soil temperatures relationships were analyzed for NWS stations that form three east-west transects to examine long-term, spatial patterns of soil temperature across the snowbelt of southern Michigan (Fig. 1). Mean monthly air temperatures, soil temperatures at 0.05 m, and air-soil temperature differences over the 40 years of record are given for the northernmost transect in Table 4. Gaylord lies in the heart of the snowbelt and has an annual snowfall that exceeds 2.5 m. Atlanta is situated to the east of the snowbelt and receives less than 2 m of annual snowfall. Snowfall totals are intermediate at East Jordan, located 19 km from Lake Michigan, and Alpena, on the shore of Lake Huron. Additionally, the frequency of deep snow cover throughout the winter is greatest at Gaylord and East Jordan, whereas Atlanta and Alpena often have years with very low snowfall totals. The modifying effect of the lakes on air temperatures is most apparent at East Jordan and least noticeable at Atlanta, due to the prevalence of westerly winds.

Inspection of the values in Table 4 reveals that temperature differences, at 0.05 m, between snowbelt and non-snowbelt sites averaged for months over the 41 years are no more than 0.6°C. The difference in soil temperature between snowbelt (East Jordan and Gaylord) and non-snowbelt (Atlanta and Alpena) sites is most pronounced during early winter (Dec and Jan) and less during late winter (Feb and March). The "insulating" effect of early winter, lake-effect snowpacks, as indicated by the air-soil temperature differences in Table 4, is especially evident at Gaylord. There, cold air masses are less effective at cooling the upper soil layers due to thick, early winter snowpacks. Thus, the soil stays warm longer into the winter and rarely reaches the low temperatures of the non-snowbelt sites (Atlanta and Alpena). Soils at East Jordan do not get as cold as non-snowbelt sites because the temperature of the coldest air masses is moderated by the relatively warm (unfrozen) lakes, and because of early winter snowpacks. Due to the presence of late-lying snow, the soil at Gaylord warms more slowly during early spring (April) than at Atlanta. Sites near the lakes (East Jordan and Alpena) also warm up slowly, due to the effect of Lakes Michigan and Huron. Most importantly, Table 4 reveals that, on average, soil temperatures at 0.05 m remain above freezing at Gaylord and East Jordan throughout the cold season, while soil temperatures at Atlanta and Alpena remain below freezing for three consecutive winter months. At deeper levels within the soil, subfreezing temperatures are rare at all sites, but snowbelt stations remain warmer than do the non-snowbelt sites. Additionally, the average minimum monthly temperatures for Atlanta (non-snowbelt interior site) is 1.7°C lower than for Gaylord (snowbelt interior site). These differences between snowbelt and non-snowbelt locations have important implications to pedologic, geomorphic, and biological processes that are effected by state changes of water in the soil.

SPATIAL PATTERNS OF SOIL FREEZING IN SOUTHERN MICHIGAN

Soil temperature calculations presented in Table 4 suggest that soils are often frozen at the 0.05 m and 0.1 m depths from December through February in northeastern lower Michigan. To the west, in the snowbelt and near the shore of Lake Michigan, soil temperatures usually remain above freezing throughout winter at all depths. Computations for the other, more southerly transects (not presented here) show similar, but less pronounced patterns. Soils in the eastern half of the central transect (Fig. 1) are usually frozen for the same three month period, but are not as cold in upper layers nor is frost as deep as in the northernmost transect. Soils in the southern transect are rarely frozen.

A surrogate for soil freezing was determined from the model's output as the number of 24 one-hour periods in which the temperature remains below -1°C (a "freeze day"). The maximum numbers of freeze days occurs in northeastern lower Michigan, where minimum soil temperatures are also lowest (Atlanta). To the west, earlier and thicker snowpacks effectively insulate the soil and inhibit freezing. In coastal areas, the moderating effects of the Great Lakes result in mean and extreme minimum winter air temperatures that are slightly higher than at interior sites. To the south, warmer air temperatures in winter lead to fewer freeze events. Consequently, soil freezing in southern Michigan is most frequent to the east of the lake-effect snowbelt, where cold air temperatures combine with thin, late winter snowpacks (Schaetzl and Isard 1991).

An important aspect of mid- and high-latitude soils is the cyclicity and magnitude of freeze-thaw events, which is related to processes such as mechanical weathering (Fraser 1959) and frost heave. Freeze/thaw cycles, defined here by -1° (lower) and 0°C (upper) temperature thresholds, are also maximal in northeastern lower Michigan where 20 or more cycles can be expected per winter season. Our data on the spatial distribution of freeze-thaw cycles (not shown here) agree very closely with that of Russell (1943), produced 50 years earlier. His map shows a maximum of freeze/thaw cycles centered on West Branch in northeastern southern Michigan, with low values near the Great Lakes.

The number of freeze/thaw cycles at 0.1 m is approximately 50-75% of the number for the 0.05 m depth. Likewise, the number of freeze days at 0.1 m are 60-80% of the number at 0.05 m. Because the soil temperatures predicted by the algorithm are 1.6°C lower than observed for the 0.05 m level during winter season (Table 2), it is likely the number of freeze/thaw cycles and freeze days calculated by the model are slightly overestimated.

CONCLUSIONS AND IMPLICATIONS

These simulations show the pronounced impact that the Great Lakes have on soil temperatures in southern Michigan, primarily through their influence on patterns of winter precipitation and air temperatures. Although similar relationships have been shown elsewhere (Bay et al. 1952; Hart and Lull 1963; Janson 1964), ours may be the first study to quantify these relationships at a regional scale. Soil freezing in winter can be important to soil formation processes such as podzolization, because the presence and type of frost can dramatically influence the amount of potential water movement through the soil (Trimble et al. 1958; Burt and Williams 1976). Because the simulations indicate that soils are least likely to be frozen where snowpacks are thickest, the spatial co-occurrence of thick snowpacks and warmer, unfrozen soils suggest possible process linkages between large pulses of infiltrating snowmelt water and (1) soil genesis, (2) groundwater recharge, and (3) release of perennial vegetation from winter dormancy.

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