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Cold Regions Science and Technology 23 (1995) 317–332

cold regions science and technology

Estimating soil temperatures and frost in the lake effect snowbelt region, Michigan, USA

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Received 13 January 1994; accepted 24 July 1994

Abstract

A physically-based computer model is developed for estimating soil temperatures at 0.05, 0.1, 0.2, and 0.5 m in sandy, forested soils, using only daily precipitation and minimum and maximum air temperatures as input data. The algorithm is evaluated by comparing its output to 490 soil temperature observations for 1990–93 at 10 sites in Michigan, USA. The mean bias of the errors of the soil temperature estimates ranges from +0.6° to −1.6°C, depending upon depth and season, thus comparing favorably to models requiring more detailed input.

The model is applied to 14 stations across lower Michigan, using 1951–91 US National Weather Service 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 lake-effect snow belt of southern Michigan and along the coast of Lake Michigan. Soil freezing is more common at non-snowbelt, interior locations. Based on simulation data, new "potential freeze day" indices are presented that correlate better with annual minimum soil temperatures and "number of days frozen" than do other, more commonly used indices of freezing not specifically formulated for soils.

1. Introduction

The temperature of soils is an important factor in determining rates of biochemical reactions that occur within them (e.g., Jansson and Berg, 1985; Marshall and Waring, 1985), and is a driving force for growth of plants and soil biota (Green et al., 1984; Andrews, 1987). Soil genesis processes, as well as land use and management, are also strongly impacted by soil temperature because moisture states and fluxes are often temperature-dependent (Linell and Kaplar, 1959; Smith et al., 1964; Goetz and Mu - pller, 1969; Burt and Williams, 1976; Schaetzl and Isard, 1991). The relationship between soil temperature and soil water movement and storage is especially important in mid-latitude and polar regions where temperatures drop below freezing for much of the year (Baker, 1971; Boyd, 1973; Schmidlin et al., 1987). For this reason, taxonomic systems incorporate soil temperature characteristics into classification schemes as high as the Great Group level in Canada (Canadian Soil Survey Committee, 1978) and the subgroup level in the USA (Soil Survey Staff, 1992).

Despite their importance, data on soil temperatures are often not readily accessible, and where they do exist, are generally limited spatially, temporally, and/or to a single depth (Wisconsin Statistical Reporting Service, 1970; McDole and Fosberg, 1974; Toy et al., 1978). Soil temperatures effectively "integrate" air temperatures over periods of time, and therefore if one

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SSDI 0165-232X(94)00020-4

knows the relationship between soil and air temperatures long-term, mean atmospheric conditions can be determined for sites where meteorologic data are lacking (Shanks, 1956). Currently, long-term mean soil temperatures are frequently estimated from air temperature or groundwater data. Progress on instantaneous soil temperature prediction from widely available atmospheric data, however, has lagged. Consequently, there exists a need to establish instantaneous relationships between soil temperatures and the atmospheric factors that are commonly monitored and distributed by weather data collection agencies (cf. Kluender et al., 1993).

Studies that relate soil and/or snow temperature to atmospheric factors can be divided into two groups based on their spatial and temporal scales: (1) site-specific investigations that focus on physical process linkages within the lower atmosphere and the upper soil profile (e.g., McKay and Thurtell, 1978; Outcalt and Hinkel, 1989), and (2) studies that establish statistical relationships between air and soil temperatures (e.g., Dimo, 1967; Toy et al., 1978; McDaniel and Munn, 1985; Kluender et al., 1993). The former types of studies typically involve numerical simulation of energy and moisture fluxes, including phase changes of soil water. Models are developed from data gathered at highly-instrumented sites where the fluxes of heat, moisture, momentum, and radiation can be monitored at frequent intervals, and snow and soil properties throughout the profile are known (e.g., McComb et al., 1992). These experiments are designed to contribute to our understanding of soil heat and moisture transfer processes at bare soil sites (e.g., Wierenga and De Wit, 1970; Lascano and Van Bavel, 1983) or where the vegetation is relatively uniform (e.g., Baker, 1971). They often have practical local applications, especially for irrigation purposes. Because these physical models require detailed measurements of atmospheric variables and soil properties (e.g., Flint and Childs, 1987), they are not intended for application at large spatial scales, and their application to "naturally vegetated" or forested sites may also be unrealistic (Shanks, 1956).

Regionally-based soil temperature studies either utilize direct measurements of soil temperatures obtained at a few widely spaced sites over relatively short time periods (e.g., Dimo, 1967), or, more typically, statistically correlate soil temperatures to air temperatures and soil characteristics/processes (e.g., McDole and Fosberg, 1974; Toy et al., 1978; Meikle and Treadway, 1981; McDaniel and Munn, 1985; Macfie, 1991; Kluender et al., 1993). One such relationship commonly used for the humid midlatitudes is: MAST = MAAT + 1°(or 2°)F (Smith et al., 1964, Smith, 1986), where MAST is mean annual soil temperature and MAAT is mean annual air temperature. This relationship has proven to be very practical for soil classification, although in some regions it is especially weak. Regions that experience deep winter snowcover, for example, are notable exceptions to this "system" (Hart and Lull, 1963; Dwyer and Hayhoe, 1985; Smith, 1986). McDole and Fosberg (1974) showed that MAST is as much as 3°C warmer than MAAT in snowy regions of Idaho.

Statistical relationships between soil and air temperature data are often established for daily, or monthly/annual mean values (e.g., Mueller, 1970; McDole and Fosberg, 1974; Toy et al., 1978; Reimer and Shaykewich, 1980). With a few exceptions (e.g., Parton and Logan, 1981, Kluender et al., 1993), efforts to predict soil temperatures at daily or shorter time scales from routinely collected (i.e., daily) atmospheric data are lacking. Consequently, most empirical air–soil temperature relationships are unable to provide insight into processes that may occur over short time scales, such as soil freeze/thaw cycles at mid-latitude continental interior locations. In addition, the insulating effect of winter snow packs in many regions make air–soil temperature relationships complex and difficult to model, especially for time periods of a week or less (Shanks, 1956; Baker, 1971).

The purpose of this study is to present and evaluate soil temperature simulations for Michigan, USA, generated by a computer algorithm that uses as input daily precipitation and minimum and maximum air temperatures collected by the US National Weather Service (NWS). This region was chosen because it experiences sub-freezing temperatures and has great variability in snow accumulations, both spatially and temporally. The algorithm computes soil temperatures at multiple depths. The simulated temperatures are verified by using measurements from 10 sites across the region, and are compared to other methods for estimating freezing and minimum soil temperatures. The model is then used to simulate soil temperatures over 41 years at 14 NWS sites across the region. Regional patterns of soil temperature are analyzed to show the modifying effect of snow thickness on air–soil temperature relationships over the 41 years. The utility of a number of simple indices for indicating frozen soil conditions is explored. Finally, the implications of the simulations to soil classification, as determined by the US Soil Conservation Service (SCS), is discussed.

2. Model construction

A computer algorithm was developed to calculate the vertical profile of temperature for coarse-textured, well-drained, forested soils. Although not a focus of this paper, the model also has a water balance component that calculates daily infiltration of water through the litter layer and into the mineral soil (Fig. 1). The model combines a Thornthwaite-based water budget model (Thornthwaite 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). In this respect, it resembles a compartment model as discussed by Kline (1973). Details of an earlier version of the model without the heat conduction component are provided elsewhere (Schaetzl and Isard, 1991).

The model is driven by daily maximum (T_{max}) and minimum (T_{min}) air temperatures (°C) and precipitation data, which were obtained by the NWS. Precipitation that falls when the mean daily temperature (T_{mean}) is ≥ 0°C is assumed to be rain; other precipitation is snow, using a 10 : 1 conversion for solid to liquid precipitation. The amount of water reaching the forest floor via stemflow and throughfall is computed from precipitation data as a function of forest type (mixed vs. deciduous), season of the year (leaf-on vs.

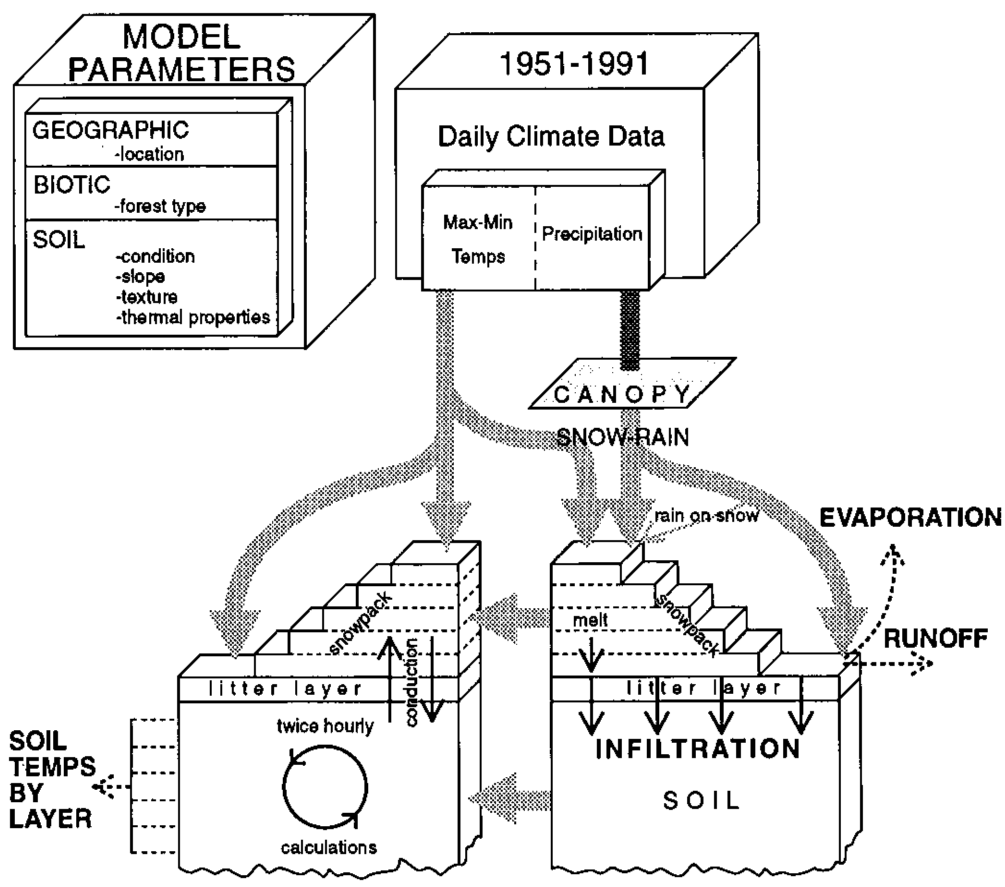

Fig. 1. Diagrammatic representation of the soil temperature and hydrology model.

leaf-off), and precipitation type (snow vs. rain). (See Schaetzl and Isard, 1991 for equations and a review of the hydrologic literature.)

Snow may be temporarily stored in a snowpack if throughfall and air temperature conditions allow. When the $T_{mean} > 0°C$ and a snowpack is present, snowmelt is calculated as a function of mean daily air temperature (T_{mean}):

$$\text{Melt (mm day}^{-1}) = 1.27T_{mean}$$

(Garstka, 1964) and is made available for storage in the litter layer. Liquid throughfall may also be stored in the forest litter layer. On days when the computed litter moisture content (composed of antecedent litter content plus melt and throughfall) exceeds the litter moisture retention capacity (10 mm), the difference may either infiltrate into the mineral soil or run off. Runoff is computed as a function of precipitation, antecedent precipitation, forest condition, and soil group classification (USDA–SCS, 1971). On days when snow cover is absent and $T_{mean} > 0°C$, evaporation of water from the litter layer is calculated as a linear function of T_{mean} using the relationship between the ratio of actual to potential evaporation and available moisture (Thornthwaite and Mather, 1955; Baier and Robertson, 1966; Willmott, 1977).

The temperature profile within the mineral soil is calculated at 30 min intervals using a finite difference formulation with nine soil layers that increase in thickness with increasing depth (0.05, 0.1, 0.3, 0.5, 1.0, 1.5, 3.5, 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 MAAT as suggested by Geiger (1965) and Smith et al. (1964). Temperatures (T_t) at the soil surface for the 30 min intervals were calculated following Parton and Logan (1981) using a truncated harmonic function of time around the T_{mean} for daytime:

$$T_t = (T_{max} - T_{min})\sin[(\pi m)/(D+1)] + T_{min}$$

and an exponential function for nighttime:

$$T_t = T_{min} + (T_{ss} - T_{min})e^{-1.81n/N}$$

D and N are day and night length (hr), respectively. T_{ss} is the soil surface temperature (°C) calculated for sunset, m is the number of hours between the current time interval and sunset, and n is the number of hours between sunset and the current time interval.

Thermal properties for the soil, litter, and snowpack are taken from Van Wijk and De Vries (1963). Thermal conductivities (mcal/cm sec °C) and volumetric heat capacities (cal/cm^3 °C) for sandy soils are 0.7 and 0.3, 4.2 and 0.5, and 5.2 and 0.7 for dry, moist and wet conditions, respectively. Litter thermal conductivities and volumetric heat capacities were assumed to be 0.6 mcal/cm sec °C and 0.6 cal/cm^3 °C. Finally, snowpack thermal conductivity and volumetric heat capacity values were set at 0.32 mcal/cm sec °C and 0.2 cal/cm^3 °C.

Site-specific input parameters are needed to operationalize the algorithm: (1) latitude is used to calculate evaporation and soil surface temperature, (2) forest type (deciduous vs. mixed) is used in throughfall calculations, (3) water retention capacity of the litter layer affects water balance calculations, and (4) SCS hydrologic soil group classification and SCS forest condition classification are needed for estimations of runoff.

3. Model verification and accuracy

A soil temperature measurement network was initiated in 1990 across southern Michigan; 14 sites are currently operational (Fig. 2, Table 1). Copper–constantan thermocouples were installed at each location at 0.05, 0.1, 0.2, and 0.5 m below the mineral soil surface. Soil temperature at each depth, snowpack thickness (mean of three readings), and time of observation are recorded bimonthly at each site. All sites are located within 17 km of a NWS station (Table 1). All 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 measurement sites) were used to evaluate the accuracy of the modelled soil temperatures.

Table 2 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. The predicted soil temperatures are more accurate for the warm season (May–October) than for the cold season (when snowpacks are usually present, cf. Reimer and Shaykewich, 1980). 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

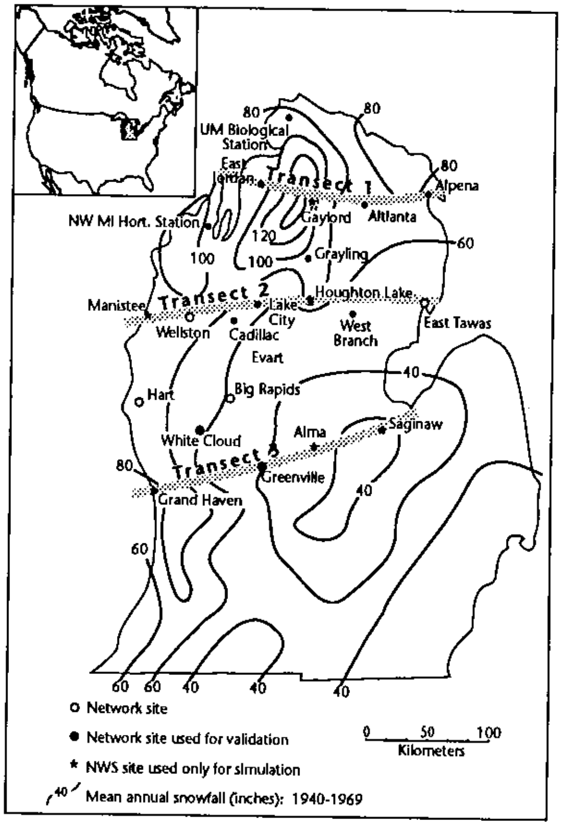

Fig. 2. Map of southern Michigan showing locations of US National Weather Service stations used in this study. The soil temperature network (monitoring) sites are indicated with a circle; open circles represent those used for model verification purposes. NWS sites that were used in the analysis of 1951–91 soil temperature trends are located with ★. 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.

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 these errors are discussed below.

Inspection of the root mean square error statistics (RMSE, Willmott, 1984) 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 2). This corresponds to a decrease in diurnal variations in soil temperature with depth. The index of agreement (*d*, Willmott, 1984) 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. If $RMSE_s$ is defined as the *systematic* error associated with misspecifications within the model (cf. Willmott, 1984), 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 such systematic causes. The systematic, or model-based, error is $\leq 5\%$ of the total error for the 0.5 m level. The large systematic errors ($\geq 50\%$ 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. It should also be noted that the model does not take latent heat into account; its omission may have contibuted to the relatively large systematic errors for the cool season.

Calculated vs. observed soil temperatures for the 0.05 and 0.5 m depths are plotted in Fig. 3a and b. The summer "warm bias" (overestimation of warm season temperatures) and the winter "cold bias" of the model are apparent at 0.05 m (Fig. 3a) but not at 0.5 m (Fig. 3b). Overestimation of soil temperatures during the warm season has been reported by others (Mueller, 1970; Vose and Swank, 1991); the error has been as high as 4°C (Hayhoe et al., 1990). Our relatively small "warm bias" likely occurs because of large differences in air temperature and radiation, especially on sunny days, between NWS instrument shelters in open areas and neighboring shaded forest sites. These differences are greater during summer than in the other seasons and produce overall cooler soil temperatures beneath the vegetative canopy than at the open, NWS site (Hayhoe et al., 1990; Balisky and Burton, 1993).

The RMSEs for the stations compare favorably with the expected errors of 1 to 3°C associated with a soil

Table 1
Characteristics of the soil temperature measurement sites

Site	Direction and distance of measurement site from NWS station	Soil series and surface texture	Elev. (m)	Vegetation at site	Slope and aspect
Pellston	same location	Rubicon sand	234	White pine, sugar maple, aspen	5% N
Atlanta	8.7 km S	Mancelona loamy sand	356	White oak, aspen, red oak, red maple	none
East Jordan	3.2 km WSW	Emmet loamy fine sand	276	Beech, red oak, sugar maple, basswood	none
Lake City	0.9 km NNW	Montcalm sand	377	Red pine	2% N
Evart	6.6 km NNW	Montcalm loamy sand	362	Sugar maple, ironwood	5% ENE
Big Rapids	11.3 km NW	Rubicon sand	324	Red pine	3% E
West Branch	5.9 km NNW	Rubicon sand	330	Red pine	2% N
Grayling	16.3 km ESE	Grayling sand	344	Jack pine, aspen	none
Wellston	same location	Rubicon sand	291	White oak, red oak, red pine, red maple	none
Greenville	7.8 km NNW	Rubicon loamy sand	262	Red pine, red oak, white ash	1% NW
NW Michigan Hort. Station	0.8 km W	Leelanau sandy loam	232	Sugar maple, white ash, ironwood	none
East Tawas	6.8 km ENE	Rousseau fine sand	186	Jack pine, pin oak	none
Hart	9.7 km NNE	Croswell sand	198	White pine, oak	none
Alpena	0.35 km S	Grayling sand	210	Red pine, black oak, jack pine	2% N

Table 2
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

temperature model recently developed by Vose and Swank (1991). In that study, however, mean daily soil temperatures for a snow-free, forested site in North Carolina were simulated using hourly air temperature inputs.

Our data support the generalization that snowpacks under forest cover accumulate and melt more slowly than do those in the open (Hoover, 1971). There is a weak association between the calculated snowpack depths and those observed at the forested sites. The algorithm often appears to melt the snowpack too rapidly on days when the T_{mean} 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 periods of melt. 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 buffering from the cold air. This shortcoming of the model is in large part due to the

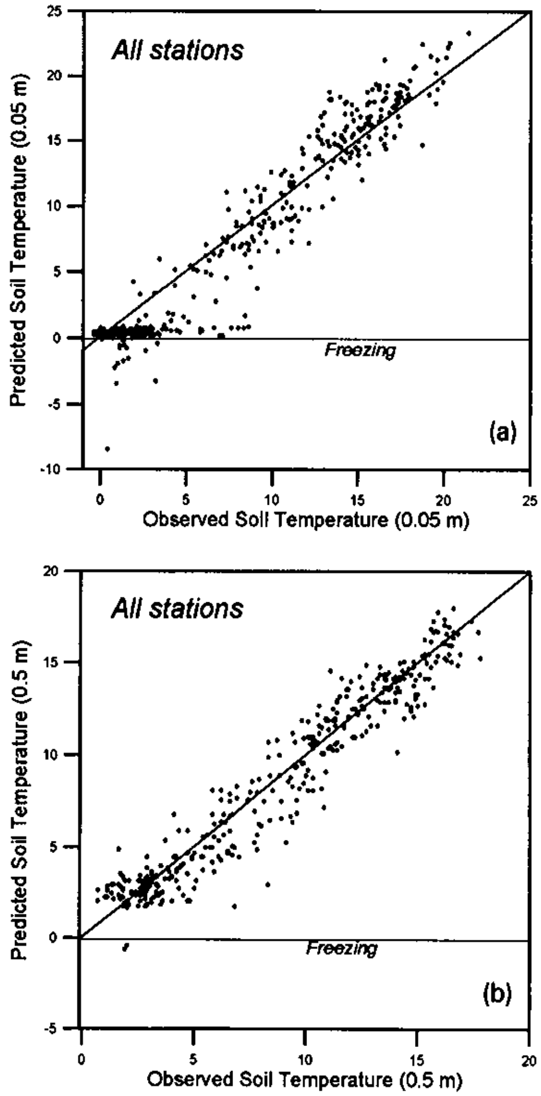

Fig. 3. Relationship between observed and predicted soil temperatures at the 0.05 m (a) and 0.5 m (b) depths for all measurement sites.

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 (Zuzel and Cox, 1975). The model employed by Hayhoe et al. (1993) illustrated the difficulties encountered in estimating snowpack thickness with limited meteorological data as inputs.

Other limitations of the snowpack component of the model include the fact that it uses a constant to convert liquid precipitation to snow and does not account for changes in snow thermal properties as the pack ages or "ripens". Varying the coefficients and formulations of the throughfall and snowmelt equations and the linear conversion of liquid precipitation to snowfall did not systematically improve the model's ability to replicate snowpack presence/absence and thicknesses.

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 coefficient, 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 sensitivity analysis on an earlier version of the model), but the changes had little effect on 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 3 shows the soil temperature model errors associated with individual stations, aggregated for all depths and seasons. The errors for the individual stations, with the exception of Atlanta, are very similar to those for the combined data set (Table 2). 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.

4. Regional soil temperature simulations

Southern Michigan's lake-effect snow belt occurs in a north–south tract approximately 40 km inland from Lake Michigan (Eichenlaub, 1970, Eichenlaub et al., 1990). Snowfall in the lake-effect region tends to accumulate earlier in winter than at sites elsewhere (Dewey, 1971).

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 (Fig. 2). Soil temperatures at the 0.05, 0.10, 0.20, and 0.5 m depths over a 41 year period (1951–1991) were simulated using NWS data from each station. Since most of the soils in this area are sandy, we feel that the model is appropriate to use for regional soil

Table 3
Soil temperature model error statistics for individual stations. Annual series averaged for all depths

Station	n	r^2	RMSE (°C)	d	MBE (°C)	$(RMSE_s/RMSE)^2$
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
Grayling	108	0.93	1.8	0.97	−0.7	0.26
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
East Tawas	20	0.94	1.3	0.96	−1.0	0.63
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

Table 4
Correlation coefficients between mean monthly air and soil temperatures ($n = 574$)

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

temperature comparisons. Model outputs include air and soil temperatures, snowpack depth, and a wide variety of freezing indices for each month or season of record. In order to characterize regional patterns of soil temperature and freezing across southern Michigan, many of these factors were averaged over the 41 year record for each station.

5. Air–soil temperature relationships

Correlation coefficients for relationships between mean monthly air and soil temperatures are provided in Table 4. Strong positive correlations ($P < 0.001$) exist 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 point to the buffering effects of the soil above. Because temperatures in deeper soil layers respond more slowly to air temperature fluctuations (a "lag effect"), they are often also 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 coefficient increases to 0.93. This lag effect on the relationship between air temperature and soil temperatures at depth was especially apparent for the "transitional" spring and autumn seasons (cf. Langholz, 1989). 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.

Table 5 illustrates that the difference between mean monthly air and soil temperatures for the 0.05 m depth are positively correlated with mean snowpack thick-

Table 5
Mean monthly values and correlation coefficients for relationships between air and soil (0.05 m) temperature differences and snowpack depth

	Oct	Nov	Dec	Jan	Feb	Mar	Apr
Mean snowpack depth ($\times 10^{-2}$ m)	0.03	2.3	17.6	42.2	44.5	71.1	26.3
Air–soil temperature difference (°C)	−0.2	−1.0	−3.8	−6.6	−6.0	−1.5	−2.4
Correlation coefficient	−0.11	−0.34	−0.56	−0.34	−0.33	−0.39	−0.58

ness (Smith et al., 1964). This relationship is strongest for late autumn and early winter, after packs in the snowbelt of southern Michigan become thick enough to insulate the soil from bitterly cold air masses that invade the region. Frequently, cold air masses follow cold fronts that bring deep lake-effect snows. Snowpacks are at their greatest depths in mid-winter, but the air–soil temperature differences are less because by that time the soil has had ample time to lose heat.

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 (Fig. 2). 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 6. 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 snowfall per annum. Annual 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.

Close inspection of the values in Table 6 reveals that soil 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 6, is especially evident at Gaylord. There, cold air masses are less effective at cooling the upper soil layers due to thick, early winter snowpacks.

Table 6
Mean monthly air temperature and (computed) soil temperatures (0.05 m depth) for a transect of stations across the snowbelt: 1951–1991 data

Site	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	

Thus, the soil stays warm longer into the winter and rarely reaches the low temperatures of the non-snowbelt sites (Atlanta and Alpena). Due to the presence of late-lying snow, soils at Gaylord warm 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 6 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 are usually below freezing for the three winter months. At deeper levels within the soil, subfreezing temperatures are rare at all sites, although snowbelt stations still remain warmer than do the non-snowbelt sites. Additionally, the average minimum monthly soil temperature 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 affected by state changes of water in the soil.

6. Evaluation of soil freezing indices for Southern Michigan

Indices of cold and soil freezing are used for a wide variety of applications including: (1) assessing winter severity, (2) predicting date of freeze for water bodies, (3) estimating snowfall and duration of snow cover, and (4) determining likelihood of soil freezing (Boyd, 1973; Bilello and Appel, 1979; Assel, 1980; Schaetzl and Isard, 1991). For practical reasons, most of these indices are formulated as a function of air temperature and consequently do not capture the modifying effect of snowcover on temperatures of substrates below. The soil temperature algorithm presented in this study calculates a number of freezing indices (described below), allowing for an evaluation of their use for estimating soil "coldness" (Table 7).

Richards (1964) defined a freezing degree day (FDD) index as the departure of the mean daily air temperature from 0°C, considering only days with temperatures below freezing. Boyd (1973) developed a seasonal degree–day index by summing FDDs and TDDs (defined as the departure of T_{mean} above 0°C). Because FDDs are always positive, they are subtracted from the index value while TDDs (also positive values) are added. Boyd's degree–day range (DDR) index is equal to the range of values (highest autumn minus lowest spring points) on the cumulative degree–day time curve for a 12 month period beginning and ending in autumn. Bilello and Appel (1979) constructed their mean degree day (MDD) index by summing the annual departures of the long-term daily mean

Table 7
Statistical relationships (r-values) between freezing indices and soil "coldness" [a]

Indicator of soil "coldness"	Assel (1980)	Boyd (1973)	Bilello and Appel (1979)	Schaetzl and Isard (1991)		This study	
	SDD	DDR	MDD	$PFD_{Sn.1,Tmx0}$	$PFD_{Sn.1,Tmn\text{-}6.7}$	$PFD_{Sn.2,Tmx0}$	$PFD_{Sn.2,Tmn\text{-}6.7}$
Min soil temp (0.05 m)	0.15	0.24	0.17	0.74	0.73	0.77	0.78
Min soil temp (0.1 m)	0.18	0.28	0.21	0.73	0.74	0.77	0.79
Min soil temp (0.2 m)	0.23	0.33	0.27	0.70	0.70	0.75	0.77
Min soil temp (0.5 m)	0.36	0.46	0.40	0.56	0.59	0.61	0.64
Freeze increments/yr (0.05 m)	0.28	0.34	0.34	0.65	0.62	0.73	0.73
Freeze increments/yr (0.1 m)	0.26	0.32	0.32	0.64	0.63	0.71	0.73
Freeze increments/yr (0.2 m)	0.24	0.29	0.30	0.60	0.62	0.66	0.69
Freeze increments/yr (0.5 m)	0.17	0.19	0.20	0.29	0.32	0.30	0.34

[a] SDD is seasonal degree–day index; DDR is degree–day range index; MDD is mean degree–day index; $PFD_{Sn.1,Tmax0}$ is potential freeze day index with 0.1 m snow depth and 0°C maximum daily temperature thresholds; $PFD_{Sn.1,Tmn\text{-}6.7}$ is potential freeze day index with 0.1 m snow depth and −6.7°C mean daily temperature thresholds; $PFD_{Sn.2,Tmx0}$ is potential freeze day index with 0.2 m snow depth and 0°C maximum daily temperature thresholds; $PFD_{Sn.2,Tmin\text{-}6.7}$ is potential freeze day index with 0.2 m snow depth and −6.7°C mean daily temperature thresholds; Min soil temp is minimum soil temperature (°C) and Freeze periods/yr are the number of 30 min increments or time steps in the model for which soil temperature is < −1°C; the number of observations (years) for these relationships is 41.

temperatures from 0°C for the period of the year when mean air temperatures are below freezing. Assel (1980) modified the MDD index for use in individual years. He begins the FDD accumulation on the first day after 1 October that the mean daily temperature falls below freezing and continues accumulating until 30 April of the following year. During this period, the seasonal degree-day (SDD) index is constrained to remain at zero or below. Schaetzl and Isard (1991) developed a more flexible potential freeze day (PFD) index. We defined PFD as the number of days during a year (1 October–30 September) in which snow cover is less than a specified depth *and* air temperature falls below a specified level. For this study four PFD indices are defined (see Table 7 footnote for details); the snowpack depth threshold is set at 0.1 and 0.2 m and the mean air temperature threshold is set at 0 (T_{max}) and −6.7°C (T_{mean}).

The soil temperature model generates two indicators of soil "coldness" that are correlated with the freezing indices described above. These are: (1) minimum soil temperature each winter for each soil depth (MST), and (2) freeze increments (FI), calculated as the number of 30 min intervals each winter for which the soil temperature is < -1°C. The indices of soil freezing described above and the soil "coldness" indicators (MST and FI) based on simulated soil temperatures were calculated for each depth and year (1951–91) for the 14 stations.

Relationships between the DDR, MDD, and SDD indices and soil "coldness" (MST and FI) are weak (Table 7). The correlations of the indices with MST increase with depth. In contrast, the correlations of the DDR, MDD, and SDD with FI decrease with depth. The reasons for these trends are not apparent. The PFD indices of Schaetzl and Isard (1991), which were explicitly developed to estimate the likelihood of soil freezing, perform better for all depths than do the DDR, MDD, and SDD, because the PDF indices account for the insulating properties of snowcover. The PFD indices are based on the generalization that loss of heat from the soil is minimal when snow cover is greater than some predetermined thickness. Schmidlin et al. (1987) used a similar index in their study of freeze–thaw cycles; freeze–thaw days (based on air temperatures) with <7.5 cm of snow cover were deemed more "effective" than were days with deeper snowpacks. When a Potential Freeze Days index is formulated with

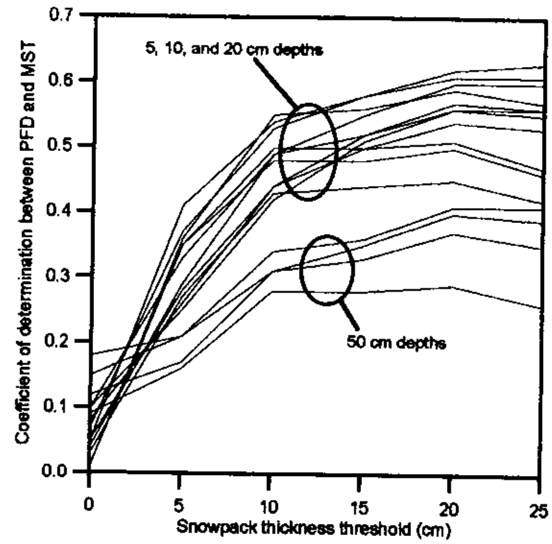

Fig. 4. Relationships between Potential Freeze Day–Minimum Soil Temperature correlations and snowpack thickness threshold.

a 0°C minimum temperature threshold but no consideration of snowpack depth, the correlations with MST (0.09, 0.12, 0.16, and 0.30 for the four depths) are similar to those for the DDR, MDD, and SDD indices in Table 7. The correlations between each of the four PFD indices and FD and MST also decrease in strength with depth, because the upper soil layers provide insulation from extreme air temperatures for lower soil layers and consequently decrease the influence of snowpack thickness during winter on soil temperatures.

An analysis of relationships between PFD indices and soil "coldness" (as represented by FD and MST values) can provide an indication of the maximum effective insulating depth for snowpacks in this region. For example, Geiger (1965) noted that the daily temperature range is damped by 50% under 7.5 cm of snow. The strength of association between minimum soil temperatures and several PFD indices that utilized a variety of air temperature and snowpack thickness thresholds was examined to find the value (thickness) beyond which additional snow has little additional insulating effect. The results for four PFD indices are plotted in Fig. 4. Increasing the snowpack depth threshold up to 0.2 m strengthens the relationship between the computed PFD indices and MST. Any further increase in snowpack depth threshold, however, decreases the ability to predict soil "coldness" using the PFD, suggesting that, in southern Michigan, increases in snow depth above 0.2 m may provide little or no additional insulation from extremely cold air temperatures, for the soil below.

7. Spatial patterns of soil freezing in Southern Michigan

Soil temperature calculations presented in Table 6 suggest that soils are frequently 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, soils usually remain unfrozen throughout winter. 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. 2) 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 (Fig. 2) are rarely frozen.

Soil freezing, measured as the number of continuous 24 hour periods in which the temperature remains below −1°C (a "freeze day"), is shown in Fig. 5a for the 14 NWS sites. The maximum numbers of freeze days occur in northeastern lower Michigan, where minimum soil temperatures were also lowest (Atlanta, Table 6). 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 (Schmidlin and Roethlisberger, 1993). 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 are related to processes such as mechanical weathering (Fraser, 1959) and frost heave (Hershfield, 1979). The latter has implications for highway construction and engineering. Freeze/thaw cycles, defined here by −1° (lower) and 0°C (upper) temperature thresholds, are also maximal in northeastern lower Michigan where 2.0 or more cycles can be expected per winter season (Fig. 5b). Our map (Fig. 5) agrees very closely with that of Russell (1943), produced

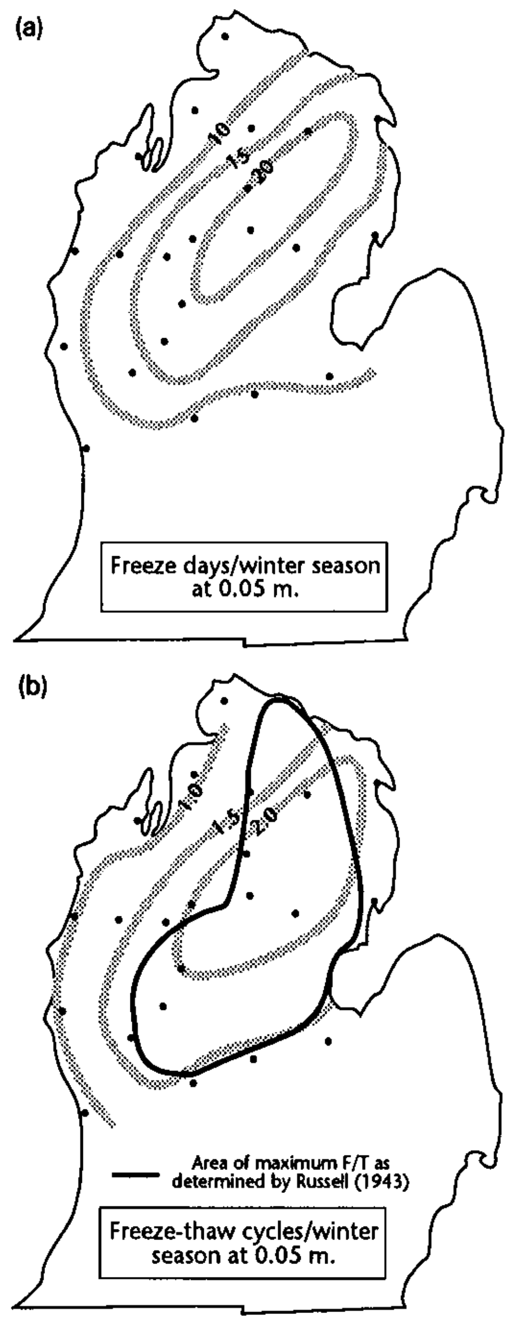

Fig. 5. Isolines of calculated (a) freeze days and (b) the number of freeze–thaw cycles for southern Michigan. Freeze days are the number of 24 hour periods in which the temperature remains below −1°C. Freeze–thaw cycles are defined by −1 (lower) and 0°C (upper) temperature thresholds. Areas enclosed by Russell's (1943) line experienced >68 freeze–thaw cycles per winter, based on air temperatures for 1914–1931.

some 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 modelled 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 those 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 the number of freeze/thaw cycles and freeze days displayed in Fig. 5 are overestimated (cf. Hayhoe et al., 1993). Notwithstanding this source of error, we believe that Fig. 5 is a good representation of the spatial patterns of freeze/thaw cycles and freeze days throughout southern Michigan.

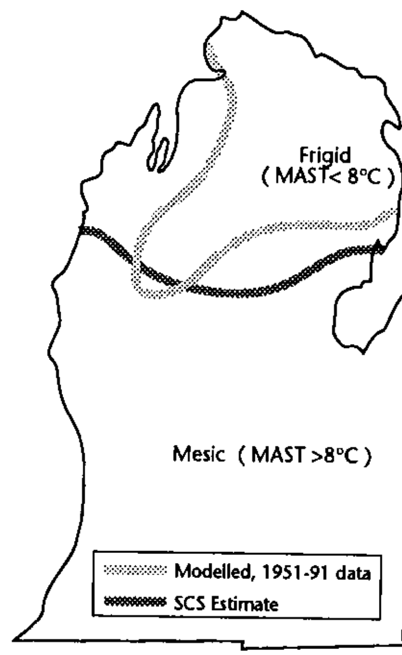

Fig. 6. Mesic–frigid soil boundary in southern Michigan soils based on a mean annual soil temperature (MAST) of 8°C at the 0.5 m depth, as currently currently used by the Soil Conservation Service, and as calculated by the soil temperature model, using data from 1951–1991.

8. Implications and conclusions

The results of 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 (Hart and Lull, 1963; Janson, 1964; Soveri and Varjo, 1977), 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 certain types of frost can dramatically influence water movement through the soil (Burt and Williams, 1976). In our study area, as in many snowy climates, large infiltration events are most common during spring snowmelt (Schaetzl and Isard, 1990). The simulations indicate that soils are least likely to freeze where snowpacks are thickest. The spatial co-occurrence of thick snowpacks and warmer, unfrozen soils suggest possible process linkages between large pulses of infiltrating water and (1) soil genesis, (2) groundwater recharge, and (3) release of perennial plants from winter dormancy.

We previously noted the strong spatial correlation between thick snowpacks and podzolic soil development (Schaetzl and Isard, 1991). The podzolization process is driven by infiltrating water containing organic acids. Our measurements and simulations, as well as the data of others (Wisconsin Statistical Reporting Service, 1970) show that soils are rarely frozen under deep snowpacks, allowing for relatively unimpeded infiltration during snowmelt. Bleak (1970) has shown that organic materials, such as forest litter, can readily decompose under snowpacks, and Jansson and Berg (1985) demonstrated the strong correlation between soil temperatures and litter decomposition. Thus, it is likely that the deep snowpacks in Michigan and elsewhere allow for one or more large, snowmelt-driven, infiltration pulses in spring, and that these pulses are a primary driving force behind soil development and potentially early spring forest growth in this region.

Soil Taxonomy (Soil Survey Staff, 1992) uses MAST to classify soils into soil temperature regimes. The boundary between mesic and frigid soil temperature regimes, as determined by the SCS, spans lower Michigan (Fig. 6). This boundary is based on a MAST of 8°C at 0.5 m. Because the bias at the 0.5 m depth for the soil temperature simulations presented in this study is only −0.1°C, it can be used to determine MAST for the study area in southern Michigan. We believe that

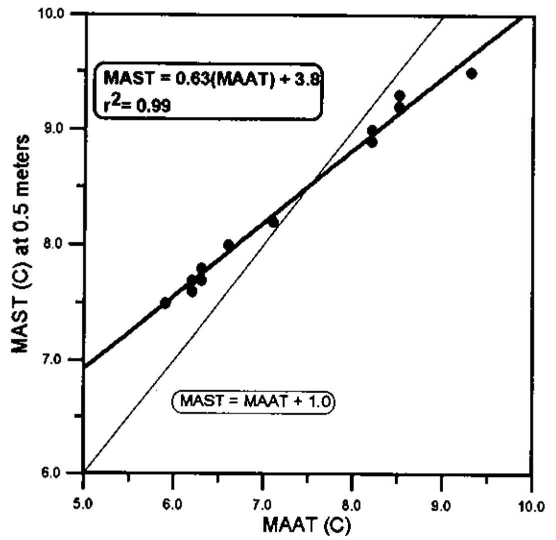

Fig. 7. Relationship between mean annual air temperature (MAAT) and mean annual soil temperature (MAST) at 0.5 m depth for 14 stations in southern Michigan for the period 1951–1991. The MAST = MAAT + 1°C line is shown for comparison.

using our modelled data is more appropriate than using the SCS's method of assuming that MAST = MAAT + 1°C, especially since this formula is known to break down in snowy climates (Smith, 1986). Fig. 7 shows the relationship between MAST calculated by our soil temperature algorithm. The thinner line represents the relationship using the SCS formula and supports Smith et al. (1964) finding that using the SCS's MAST equation for midlatitude sites often creates an error of ≈ 0.5°C. Fig. 6 also displays the mesic–frigid boundary using the calculations of MAST from the soil temperature model. The boundary proposed (at 8°C) in this study and the SCS boundary are very similar in the eastern two-thirds of the state. However, the proposed boundary curves north at approximately 50 km inland from Lake Michigan and parallels the shoreline for approximately 150 km. In contrast, the SCS boundary, which does not account for the full effect of lake effect snowfall on MAST, closely follows the 44° parallel of latitude. Consequently, our results suggest that soils in northwestern southern Michigan are incorrectly assigned by the SCS system to the frigid soil temperature regime.

Acknowledgements

This study would not have been possible without data voluntarily collected by Ray Bremer, Gary Campbell, Rod Cortright, Bill Dunn, Gary and June Ginther, Erik Johnson, Bill Klein, Bob Vande Kopple, Joe Kutkuhn, Hugo McGee, Rick Nielson, Chris Norcross, Fred Prichard, John Reedstrom, Phil Shaw, Kathy Walicki, and Tom Williams. The following organizations have kindly contributed time and personnel to the project: the USDA–SCS, the National Forest Service, the University of Michigan Biological Station, the Cooperative Extension Service and Agricultural Experiment Station of Michigan State University, the National Weather Service and the Northwest Michigan Horticultural Research Station. Special appreciation is due Fred Nurnberger and the staff at the Michigan State Climatologist's office. The figures were drafted by the Center for Cartographic Research and Spatial Analysis, Department of Geography, Michigan State University.

References

Andrews, C.J., 1987. Low-temperature stress in field and forage crop production — an overview. Can. J. Plant Sci., 67: 1121–1133.

Assel, R.A., 1980. Maximum freezing degree–days as a winter severity index. Mon. Weather Rev., 108: 1440–1445.

Baier, W. and Robertson, G.W., 1966. A new versatile soil moisture budget. Can. J. Plant Sci., 46: 299–315.

Baker, D.G., 1971. Snow cover and winter soil temperatures at St. Paul, Minnesota. Water Resources Res. Center, Univ. of Minnesota. US Dept. of Interior, Bulletin 37. 24 pp.

Balisky, A.C. and Burton, P.J., 1993. Distinction of soil thermal regimes under various experimental vegetation covers. Can. J. Soil Sci., 73: 411–420.

Bilello, M.A. and Appel, G.C., 1979. Relationships between January temperatures and the winter regime in Germany. Cold Reg. Sci. Technol., 1: 17–27.

Bleak, A.T., 1970. Disappearance of plant material under a winter snow cover. Ecology, 51: 915–917.

Boyd, D.W., 1973. Normal freezing and thawing degree–days for Canada, 1931–60. Atmospheric Environment Service, CLI 4-73. Downsview, 38 pp.

Burt, T.P. and Williams, P.J., 1976. Hydraulic conductivity in frozen soils. Earth Surf. Process., 1: 349–360.

Canadian Soil Survey Committee, 1978. The Canadian system of soil classification. Canadian Department of Agriculture Publication 1646, Ottawa.

Carslaw, J.S. and Jaeger, J.C., 1959. Conduction of Heat in Solids. Oxford Univ. Press, New York.

Dewey, K.R., 1971. The spatial distribution of lake-effect snowfall within the vicinity of Lake Michigan. Trans. Illinois St. Acad. Sci., 64: 177–187.

Dimo, V.N., 1967. Soil and air temperatures in the soil–climatic regions of the USSR. Sov. Soil Sci.: 1641–1654.

Dwyer, L.M. and Hayhoe, H.N., 1985. Comparison of observations and macroclimatic model estimates of monthly winter soil temperatures at Ottawa. Can. J. Soil Sci., 65: 109–122.

Eichenlaub, V.L., 1970. Lake effect snowfall to the lee of the Great Lakes: its role in Michigan. Bull. Am. Meteorol. Soc., 51: 403–412.

Eichenlaub, V.L., Harman, J.R., Nurnberger, F.V. and Stolle, H.J., 1990. The Climatic Atlas of Michigan. Univ. Notre Dame Press, Notre Dame, IN, 165 pp.

Flint, A.L. and Childs, S.W., 1987. Field procedure for estimating soil thermal environments. Soil Sci. Soc. Am. J., 51: 1326–1331.

Fraser, J.K., 1959. Freeze–thaw frequencies and mechanical weathering in Canada. Arctic, 12: 40–53.

Garstka, W.U., 1964. Snow and snow survey. In: V.T. Chow (Editor), Handbook of Applied Hydrology. McGraw-Hill, New York, pp. 10-1–10-57.

Geiger, R., 1965. The Climate Near the Ground. Harvard Univ. Press, Cambridge, MA, 611 pp.

Goetz, R. and Müller, S., 1969. Relationships between the daily temperature wave and the development of the natural soil profile. In: Effects of Temperature and Heat on Engineering Behavior of Soils. Highway Res. Board Spec. Rept. 103., Publ. 1641. Proc. of an Int. Conf., Washington, DC, pp. 51–65.

Green, F.W.H., Harding, R.J. and Oliver, H.R., 1984. The relationship of soil temperature to vegetation height. J. Climatol., 4: 229–240.

Hart, G. and Lull, H.W., 1963. Some relationships among air, snow, and soil temperatures and soil frost. United States Forest Service Research Note NE-3, 4 pp.

Hayhoe, H.N., Pelletier, R.G. and Van Vliet, L.J.P., 1993. Estimation of snowmelt runoff in the Peace River region using a soil moisture budget. Can. J. Soil Sci., 73: 489–501.

Hayhoe, H.N., Tarnocai, C. and Dwyer, L.M., 1990. Soil management and vegetation effects on measured and estimated soil thermal regimes in Canada. Can. J. Soil Sci., 70: 61–71.

Hershfield, D.M., 1979. Freeze–thaw cycles, potholes and the winter of 1977–78. J. Appl. Metr., 18: 1003–1007.

Hoover, M.D., 1971. Snow interception and redistribution in the forest. In: Monke, E.J. (Editor), Biological Effects in the Hydrological Cycle. Proc. 3rd Intl. Seminar for Hydrology Professors, Purdue Univ., West Lafayette, IN, pp. 114–122.

Janson, L.-E., 1964. Frost penetration in sandy soil. Transactions of the Royal Institute of Technology, Stockholm, No. 231, 165 pp.

Jansson, P.-E. and Berg, B., 1985. Temporal variation of litter decomposition in relation to simulated soil climate. Long-term decomposition in a Scots pine forest, V. Can. J. Bot., 63: 1008–1016.

Kline, J.R., 1973. Mathematical simulation of soil–plant relationships and soil genesis. Soil Sci., 115: 240–249.

Kluender, R.A., Thompson, L.C. and Steigerwald, D.M., 1993. A conceptual model for predicting soil temperatures. Soil Sci., 156: 10–19.

Langholz, H., 1989. A simple model for predicting daily mean soil temperatures. J. Agron. Crop Sci., 163: 312–318.

Lascano, R.J. and Van Bavel, C.H.M., 1983. Experimental verification of a model to predict soil moisture and temperature profiles. Soil Sci. Soc. Am. J., 47: 441–448.

Linell, K.A. and Kaplar, C.W., 1959. The factor of soil and material type in frost action. Highway Res. Board Bull., 225: 81–128.

Macfie, T.G., 1991. Estimating mean daily soil temperatures using sparse regional long-term air temperature data to assess periods of biologically active reducing conditions. M.Sc. Thesis, Cornell Univ.

Marshall, J.D. and Waring, R.H., 1985. Predicting fine root production and turnover by monitoring root starch and soil temperature. Can. J. For. Res., 15: 791–800.

McComb, T.J.L., Rimmer, A.B., Rodgers, M.L.B., Turver, K.E. and Vickers, A.F., 1992. A mathematical model for the prediction of temperature in a dry snow layer. Cold Reg. Sci. Technol., 20: 247–259.

McDaniel, P.A. and Munn, L.C., 1985. Effect of temperature on organic carbon — temperature relationships in Mollisols and Aridisols. Soil Sci. Soc. Am. J., 49: 1486–1489.

McDole, R.E. and Fosberg, M.S., 1974. Soil temperatures in selected Southeastern Idaho soils: I. Evaluation of sampling techniques and classification of soils. Soil Sci. Soc. Am. Proc., 38: 480–486.

McKay, D.C. and Thurtell, G.W., 1978. Measurements of the energy fluxes involved in the energy budget of a snow cover. J. Appl. Meteorol., 17: 339–349.

Meikle, R.W. and Treadway, T.R., 1981. A mathematical model for estimating soil temperatures in Canada. Soil Sci., 131: 320–326.

Mueller, O.P., 1970. Soil temperature regimes in a forested area of the northern Rockies. Soil Sci., 109: 40–47.

Outcalt, S.I. and Hinkel, K.M., 1989. Night-frost modulation of near-surface soil–water ion concentration and thermal fields. Phys. Geogr., 10: 336–348.

Parton, W.J. and Logan, J.A., 1981. A model for diurnal variation in soil and air temperature. Agric. Meteorol., 23: 205–216.

Reimer, A. and Shaykewich, C.F., 1980. Estimation of Manitoba soil temperatures from atmospheric meteorological measurements. Can. J. Soil Sci., 60: 299–309.

Richards, T.L., 1964. The meteorological aspects of ice cover on the Great Lakes. Mon. Weather Rev., 92: 297–302.

Russell, R.J., 1943. Freeze-and-thaw frequencies in the United States. Trans. Am. Geophys. Union, 24: 125–133.

Schaetzl, R.J. and Isard, S.A., 1990. Comparing "warm season" and "snowmelt" pedogenesis in Spodosols. In: J.M. Kimble and R.D. Yeck (Editors), Proc. 5th Int. Soil Correlation Meeting USDA–SCS, Lincoln, NE, pp. 303–318.

Schaetzl, R.J. and Isard, S.A., 1991. The distribution of Spodosol soils in southern Michigan: A climatic interpretation. Ann. Assoc. Am. Geogr., 81: 425–442.

Schmidlin, T.W., Dethier, B.E., and Eggleston, K.L., 1987. Freeze–thaw days in the northeastern United States. J. Clim. Appl. Meteorol, 26: 142–155.

Schmidlin, T.W. and Roethlisberger, R.A., 1993. Winter sub-freezing periods and significant thaws in the boreal forest region of central North America. Arctic, 46: 359–364.

Shanks, R.E., 1956. Altitudinal and microclimatic relationships of soil temperature under natural vegetation. Ecology, 37: 1–7.

Smith, G.D., 1986. The Guy Smith interviews: Rationale for concepts in Soil Taxonomy. Soil Management Support Services Technical Monograph 11 (T.R. Forbes, Editor). Cornell University, Ithaca, NY, 259 pp.

Smith, G.D., Newhall, F., Robinson, L.H. and Swanson, D., 1964. Soil-temperature regimes — their characteristics and probability. USDA–SCS Tech. Pub. 144, 14 pp.

Soil Survey Staff, 1992. Keys to Soil Taxonomy. 5th ed. SMSS Technical Monograph No. 19. Pocahontas Press, Blacksburg, VA.

Soveri, J. and Varjo, M., 1977. Roudan muodostumisesta ja esiintymisestä Suomessa vuosina 1955–1975. Publications of the Water Research Institute, Vesihallitus–Natl. Board of Waters, Finland, 66 pp.

Thornthwaite, C.W. and Mather, J.R., 1955. The Water Balance. Publications in Climatology, 8. Drexel Institute of Technology, Laboratory of Climatology, Seabrook, NJ, 86 pp.

Toy, T.J., Kuhaida, A.J. and Munson, B.E., 1978. The prediction of mean monthly soil temperature from mean monthly air temperature. Soil Sci., 126: 181–189.

USDA–SCS, 1971. National Engineering Handbook. Sect. 4: Hydrology. U.S. Government Printing Office, Washington, DC.

Van Wijk, W.R. and De Vries, D.A., 1963. Periodic temperature variations in homogeneous soil. In: W.R. van Wijk (Editor), Physics of Plant Environment. North-Holland, Amsterdam.

Vose, J.M. and Swank, W.T., 1991. A soil temperature model for closed canopied forest stands. U.S. For. Serv. Res. Paper, SE-281, 11 pp.

Wierenga, P.J. and De Wit, C.T., 1970. Simulation of heat transfer in soils. Soil Sci. Soc. Am. Proc., 34: 845–848.

Willmott, C.J., 1977. Watbug: A FORTRAN IV Algorithm for Calculating the Climatic Water Budget. Publications in Climatology, 30. C.W. Thornthwaite Assoc., Laboratory of Climatology, Elmer, NJ, 55 pp.

Willmott, C.J., 1984. On the evaluation of model performance in physical geography. In: G.L. Gaile and C.J. Willmott (Editors), Spatial Statistics and Models. Reidel, Boston, pp. MA, pp. 443–460.

Wisconsin Statistical Reporting Service, 1970. Snow and Frost in Wisconsin. USDA and Wisconsin Department of Agriculture Miscellaneous Publication, 28 pp.

Zuzel, J.F. and Cox, L.M., 1975. Relative importance of meteorological variables in snowmelt. Water Resour. Res., 11: 174–176.