Modeling Soil Temperatures and the Mesic-Frigid Boundary in the Central Great Lakes Region, 1951–2000

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ABSTRACT

Understanding the spatial and temporal variation in soil temperatures is important to classification, land use, and management. To that end, mean annual soil temperature (MAST) data for Wisconsin and Michigan were modeled to (i) determine the effects of the Great Lakes and their snowbelts on soil temperatures, and (ii) better estimate the location of the boundary between the mesic and frigid soil temperature regimes in this region. The location of the mesic-frigid (M-F) line is particularly difficult to determine where east-west gradients in air temperature cross north-south trends in snowfall due to Lake Michigan. Additionally, the soil temperature regime of several Great Lakes' peninsulas near the M-F line is in question. To determine the accuracy of our soil temperature model, soil temperature data output from it were compared with data derived from thermocouples implanted in soils at 39 sites in northern Michigan that had been collecting data several times daily for more than 6 yr. Error statistics for the model show that it has essentially no mean bias when examined on an annual basis or for winter, and only a bias of 0.1°C for the warm season. The M-F line in Wisconsin and Michigan is slightly north of most previously estimated locations, and is strongly influenced by the snowbelt in southern Michigan. Soils in deep snow areas stay warmer in winter than do soils inland, increasing their MAST and forcing the M-F line north of where air temperatures alone might have placed it. Lake-effect areas also stay cold longer into the spring season, and cool down more slowly in fall. Soil temperatures in these areas are, therefore, more moderated on an annual basis, as indicated by coefficients of variation.

Temperature is an ephemeral and constantly changing soil characteristic. Despite this, temperature data are invaluable in soil classification, necessary to understand soil genesis, useful in optimizing land use, and a primary criterion in mapping (Beckel, 1957; Smith, 1986; Ping, 1987; Larsen et al., 1988; Berry and Radke, 1995; Johnsson and Lundin, 1991; Schaetzl and Tomczak, 2002). Soil temperature data are valuable in mapping because they often are a major criterion used to determine the boundary between Major Land Resource Areas (MLRAs) (Soil Survey Staff, 1981), and soil series are defined as being unique to one soil temperature regime. In the midwest in particular, several MLRA boundaries, in theory, are designed to separate mesic (MAST > 8°C) from frigid (MAST < 8°C) soils, or at least parallel the M-F boundary. In Michigan, where north-south trending Lake Michigan and its snowbelt run counter to east-west trending climate isotherms, the location of the M-F line is particularly obscure (Isard and Schaetzl, 1995; Mokma and Sprecher, 1995). In Wisconsin, the impact of Lakes Michigan and Superior on soil temperatures inland has been little studied.

That snowbelts dramatically impact soil temperatures was recognized by the scientists who developed Soil Taxonomy. For example, Guy Smith (Smith, 1986, pl134) noted this in his "interviews":

> There is no question that the mean annual soil temperature rises with the thickness of the snow mantle that insulates the soil during the cold season. . . . In these snow belts it is doubtful that the soil ever freezes to depths of more than a few centimeters and once the snow has accumulated it is doubtful that there is any frost in the soil whatever.
>
> (G. Smith, 1986, p. 134)

The insulating effect of snowpacks on soil temperatures has been confirmed by empirical research (Geiger, 1965; Isard and Schaetzl, 1995; Schaetzl and Tomczak, 2002). Smith also realized that snowbelts confound the "traditional" practice (Soil Survey Staff, 1999) of adding ≈2°F (1°C) to the mean annual air temperature (MAAT) to arrive at MAST (Smith, 1986). This traditional method of estimating MAST was developed out of necessity because long-term records of air temperature existed for most regions, whereas soil temperature data were scarce. A problem arises, however, because the MAST = MAAT + 1°C equation breaks down in many midlatitude areas where MAST is 2 or 3°C higher than MAAT (Soil Survey Staff, 1999) and is especially problematic in snowbelt areas (Isard and Schaetzl, 1995).

A tempting alternative to the simple linear relationship between MAAT and MAST discussed above is to measure soil temperatures directly. Most soil temperature records, however, are short and have copious missing data. However, even if measured consistently over longer periods of time, soil temperature data cannot simply be averaged and assumed to be an accurate reflection of MAST, because the period of measurement may have been abnormally warm or cool, wet/snowy or dry. Thus, we argue that the best way to estimate MAST is to use a computer model that can accurately establish the complex relationship between soil and air temperatures, and then, providing that long-term, that is, ≥30 yr, of air temperature data are available, run the model. In this paper we adopted just such a modeling approach to estimate the long-term soil temperatures, and in so doing, the M-F boundary in the central Great Lakes region, where snow cover presents a complicating factor. Previous work has clearly established that soils in the Great Lakes' snowbelt areas are warmer than other soils

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Abbreviations: CV, coefficient of variation; MAAT, mean annual air temperature; MAST, mean annual soil temperature; MBE, mean biased error; M-F, mesic-frigid; MLRA, major land resource area; NRCS, Natural Resources Conservation Service; NWS, National Weather Service; PE, potential evapotranspiration; RMSE, root mean squared error; SCS, Soil Conservation Service; UP, upper peninsula.

at the same latitude (Isard and Schaetzl, 1995; Schaetzl and Tomczak, 2002). The purpose of this study, therefore, is to present modeled soil temperature data for the central Great Lakes region (Michigan and Wisconsin), which can then be used to determine the location and spatiotemporal variability of the M-F boundary in this region.

MATERIALS AND METHODS

General Principles

In this study, air temperature and precipitation data from National Weather Service (NWS) stations (Fig. 1) were input into a physically based computer model that calculates soil temperatures for various depths. The model was developed and calibrated using soil temperature measurements collected over several years at a number of snowbelt and nonsnowbelt sites in northern Michigan (Fig. 1); initial findings from this study have been published (Isard and Schaetzl, 1993, 1995; Schaetzl and Isard, 1996). To evaluate the performance of the model over a wider area and extended time period, the soil temperature measurement network was expanded to include 39 sites throughout northern Michigan. Predicted (modeled) soil temperatures, generated using weather data from corresponding NWS stations, were compared with the observed soil temperature measurements from the network. The results of this second analysis indicated that the predicted soil temperatures were reasonably accurate and nonbiased. Finally, a 50-yr record of air temperatures from over 200 NWS stations across Michigan and Wisconsin was used to calculate temperatures of well-drained soils across the same region. The results of this analysis are mapped and evaluated in this study. Details of the methodology, as well as model accuracy and bias, are presented below.

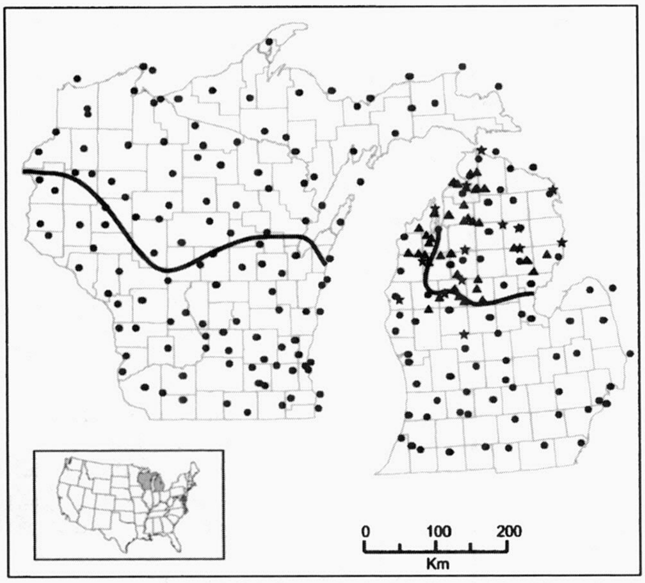

Fig. 1. Locations of data stations (three types) used in this study. 1. Stations where soil temperatures were recorded twice monthly from 1990 to 1994 (stars). Data from these stations were initially used to develop the soil temperature model. 2. Stations where soil temperatures were recorded 12 times daily 1997 to 2000 (22 401 observations total) (triangles). Data from these stations (open squares) were used to determine the accuracy of the soil temperature model. 3. National Weather Service stations in Wisconsin and Michigan used as inputs for air temperature and precipitation (dots). The best estimate of the mesic-frigid (M-F) line in the central Great Lakes region, as determined by our study, is indicated.

Data

National Weather Service daily maximum and minimum air temperatures and precipitation data from 218 stations in Wisconsin and Michigan were used as inputs to the model. The data span the time frame from 1951–2000, inclusive. The stations range in latitude from 41.7° (Morenci, WI) to 47.2°N (Houghton, MI), comprising a broad north-south transect of approximately 600 km (Fig. 1). Because the 1951–2000 data for most stations, obtained from the National Climatic Data Center in Asheville, NC, contained occasional (<10%, by our estimation) missing values, we developed a "buddy" system for estimating and filling in missing data (Schaetzl and Isard, 1996). Each primary NWS station was assigned a maximum of six buddy stations, which were usually within 50 km of their respective primary station. When a primary station had a missing value, we substituted the value from its nearest buddy. If this buddy also lacked a measurement of the climate factor for that date, we used data from the next nearest buddy, and if necessary, the third buddy, etc. Eventually, we developed a complete data set for the 218 NWS stations in the study area. The buddy system was important to this research because it allowed us to develop and use a much larger and richer data set than would have been possible if we had been restricted to the raw NWS data.

In the initial development of the model, discussed more thoroughly elsewhere (Isard and Schaetzl, 1995), real-time soil temperature data, that is, "ground truth," were collected from 14 sites throughout northern lower Michigan between 1990 and 1994 (Fig. 1). These data (490 total observations) were collected twice monthly by volunteers; automated data collection protocols were not operational at this time. Characteristics of the sites are provided in Isard and Schaetzl (1995); all are located in well-drained soils, under mature or nearly mature, broadleaf or mixed broadleaf-coniferous forest. These data were used to refine the model; results are reported in Isard and Schaetzl (1995).

A second, richer set of data was also used to expand the study of soil temperatures. Beginning in 1997, an automated network of soil temperature stations was established throughout northern Michigan; this network would eventually grow to include 39 stations (Fig. 1). Siting preference was given to areas within 50 km, either way, of the M-F boundary in Michigan, as inferred by the NRCS (Fig. 2A). Areas that NRCS personnel thought were particularly problematic with regard to soil temperature were given additional attention; an extra station or two was sited there. Most of these sites had one of two characteristics: (1) they were located near the inferred M-F line but on a nearby large upland while the other is on a well-drained plain, or (2) they were just south of the M-F line (based on earlier data) but NRCS personnel believed the site to be frigid.

Initially there were 25 stations in 1997. The network expanded by three to five each year thereafter, as data accrued, which allowed us to identify "key" and "problematic" soil temperature areas. At each site, a representative, upland, forested location on a slope of <5% was first identified and a weatherproof, copper-constantan thermocouple installed at 50 cm by implacing it ≈15 cm horizontally into an undisturbed soil profile in the face of a small pit. The 50-cm depth was chosen because it is traditionally viewed as the depth below which diurnal temperature fluctuations are damped out (Smith et al., 1964). The thermocouple was connected by cable to a small data logger, which remained concealed on the soil surface. We initially used two different models of StowAway Tidbit XT Temperature Loggers (Onset Computer Corp., Pocasset, MA), but as the technology improved we changed to HOBO Pro Data Loggers, also from Onset Corporation. The technology change represented an upgrade to a more accurate

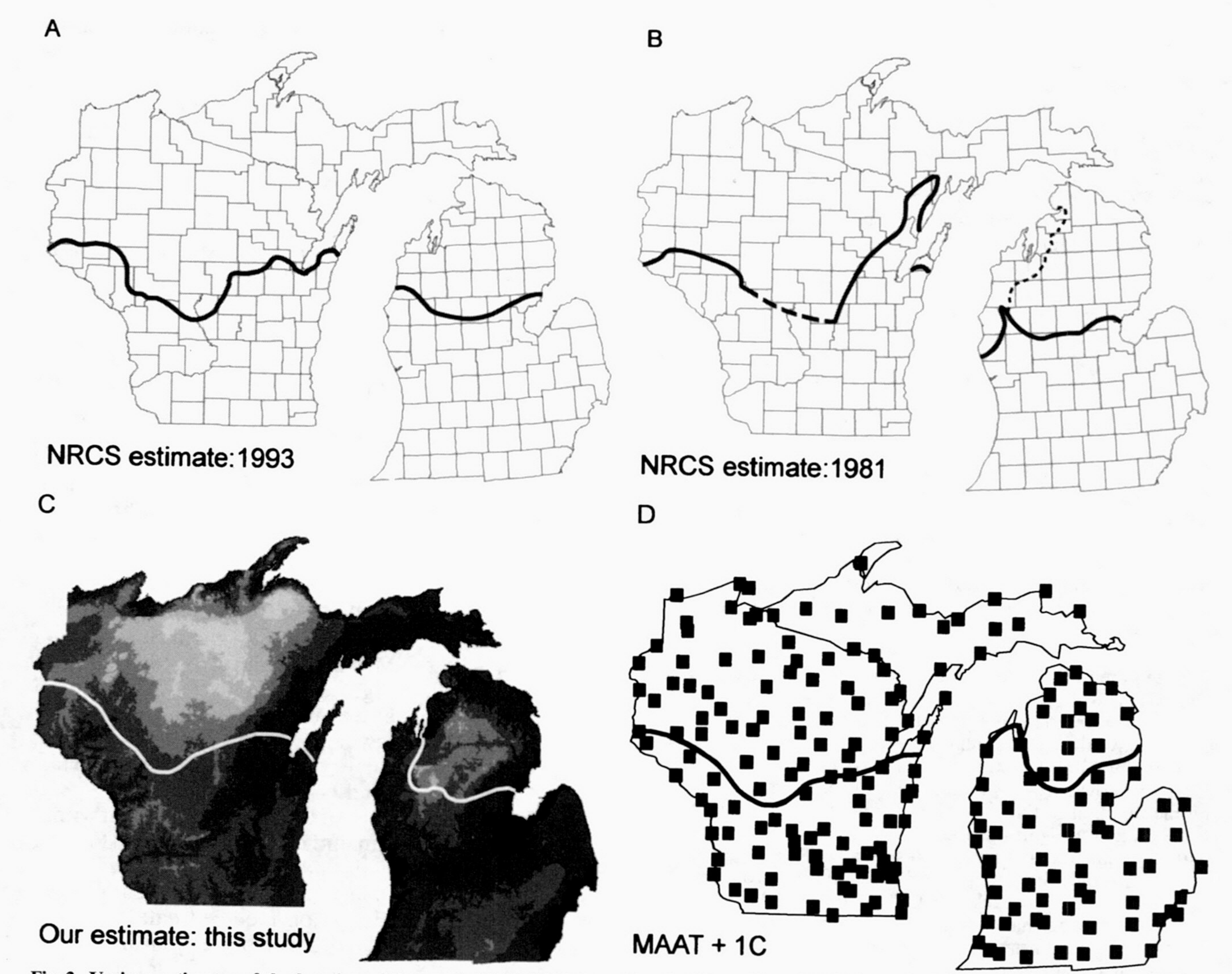

Fig. 2. Various estimates of the location of the mesic-frigid (M-F) boundary in Wisconsin and Michigan. A. Based on a USDA-NRCS map of the soil temperature regimes of the United States (USDA National Cartography and Geographic Information Systems Center 1993). B. Based on interpretations of the MLRA map of the USDA-NRCS, on-line at http://www.essc.psu.edu/soil_info/soil_lrr/. The dotted line in Wisconsin cuts across the center of a MLRA that is being mapped with both mesic and frigid soils. The dotted line in Michigan shows where, based on current soil mapping operations, the M-F line is assumed to exist by NRCS personnel; this line is not shown on published NRCS maps, which are, in turn, probably based on Agricultural Handbook 296 (USDA-SCS 1981). C. Mean location of the M-F line, on a digital elevation model (DEM) base, based on model output, using climatic data for the period 1951–2000 (see Fig. 3 for details). DEM class interval = 50 m. D. Traditionally, mean annual soil temperature (MAST) is often assumed to be equal to mean annual air temperature (MAAT) + 1°C. This map shows the location of the 8°C [MAAT plus 1°C] line, as well as the locations of our 218 NWS stations.

model with more data storage capability and a longer battery life. The HOBO Pro Data Logger has an accuracy of ± 0.2° within the temperature range encountered in the field; comparable accuracy information for the StowAway Loggers is ± 0.4°. Each logger was set to record the soil temperature at 2-h intervals. Data were downloaded annually in the field to a laptop computer. Missing data due to battery failure or vandalism were common but gaps in the data set did not greatly affect the research because we were only using this data set to establish the accuracy of the model for this wider area. These actual, instantaneous soil temperature data, totaling 22 401 observations from 1997 through the end of 2000, were compared with output from the soil temperature model as a means of establishing its accuracy and bias; these soil temperature data that were thought to be a better yardstick against which to compare the model because they were viewed as being more accurate and from a wider variety of soils than those used in our earlier study (Isard and Schaetzl, 1995).

The Model

The model functions as follows. Vertical profiles of water and temperature are calculated for well-drained, forested soils using a modified form of a soil water and temperature algorithm developed earlier (Schaetzl and Isard, 1991, 1996; Isard and Schaetzl, 1993, 1995). The model uses a Newhall-based, water budget component (Van Wambeke et al., 1986), combined with a Soil Conservation Service (SCS) snowmelt model (USDA-SCS, 1971) and a one-dimensional heat conduction equation (Carslaw and Jaeger, 1959). The model is formulated with twenty, 5-cm thick soil layers, five additional soil layers (that increase in thickness with increasing depth), one litter layer, and up to ten snowpack layers (Schaetzl and Isard, 1996). It uses soil hydrologic and thermal properties that are consistent for coarse-textured (coarse-loamy and coarser) soils, which dominate large parts of the region. Modified from Van Wambeke et al. (1986), the model simulates the progression of a wetting front into the soil. The model assumes homogeneous, piston-like percolation of wetting fronts, although finger flow is common in coarse-textured soils (e.g., Price and Bauer, 1984; Kung, 1990). At this scale of analysis the model provides a representation of average fluxes of soil water over periods from days to years. We generated hydrologic data for the soils because it influences their thermal conductivities; we do not explicitly use soil wetness or infiltration data in this study. The state (liquid vs. solid) and amount of water reaching the forest floor via stemflow and throughfall is calculated as a function of air temperature, precipitation amount, and various forest hydrology equations, specific to the time of year and forest type (Schaetzl and Isard, 1996). The water is stored in a snowpack when throughfall and air temperature conditions allow. Snowmelt is calculated as a function of air temperature, and meltwater is made available for storage in the litter and/or soil layers. Liquid throughfall can also be stored in the forest litter. The water storage capacity of the litter and soil layers—necessary parameters to run the model—are specified elsewhere (Schaetzl and Isard, 1996) and are based on field data. Thornthwaite's formula for potential evapotranspiration (PE) was used to determine the daily amount of water removed from the litter and soil layers (Thornthwaite and Mather, 1955). Water stored in the litter and uppermost soil layer is used in the PE calculation. When PE exceeds the amount of water stored in the litter and uppermost soil layer, the excess water is removed from lower layers following the procedure suggested by Van Wambeke et al. (1986). Water is sequentially withdrawn from the layers by assuming a linear relationship between the ratio of water removal to PE and available water (Baier and Robertson, 1966).

Temperature in the lowest (7 to 15 m deep) soil layer is held at 2°C above the mean annual air temperature, as suggested by Geiger (1965) and Smith et al. (1964). Temperatures at the litter or snow surface are calculated using a truncated harmonic function of time for daytime and an exponential function of time for nighttime, following Parton and Logan (1981), with daily maximum temperatures set at 1400 h Central Standard Time, and daily minima at dawn. Thermal properties for soil, litter, and snowpack are taken from van Wijk and de Vries (1963). Thermal conductivities and volumetric heat capacities for the soil layers are specified as a function of soil water (Lowrey and Lowrey, 1989). A finite difference formulation is used to calculate the temperature profile within the mineral soil at 20-min intervals.

Modeled soil temperature data were plotted on maps, and isolines drawn using ArcGIS (ESRI, Redlands, CA) and Surfer (Golden Software, Golden, CO) software packages. Routine descriptive statistics were run on the data to determine various indicators of annual variation in soil temperature, and annual soil temperature extremes, across the region.

RESULTS AND DISCUSSION

Output data from the soil temperature model were compared with the 22 401 soil temperature observations, taken at the 50-cm depth from 39 different sites, from 1997 through the end of 2000, to establish accuracy and bias. Error statistics, summarized as root mean squared error (RMSE) and mean biased error (MBE) in the soil temperature simulations, indicate the exceptional accuracy of the model (Table 1). Although there exists a small "cold" bias in summer, that is, the model predicted the soils to be, on average, 0.1°C colder in summer than they actually are, the key indicator is the absence of a bias in the annual series, and in winter. Thus, we concluded that, for our purposes, the model's predictions of MAST are almost without bias, and fall within acceptable ranges of error, for soils in the central Great Lakes region.

Table 1. Error statistics for the soil temperature model: model output vs. actual soil temperatures (at 50 cm depth) measured at 39 Michigan locations.

Period of measurement	Number of observations	Root mean squared error†	Mean biased error†
		°C	
Entire period (annual)	22 401	1.5	0.0
Warm season (May-Oct)	11 142	1.9	−0.1
Cold season (Nov-Apr)	11 259	1.1	0.0

† A positive error or bias indicates that the model predicted warmer temperatures than actually existed in the field. Negative errors or biases indicate the opposite.

The M-F line, initially established to separate areas which could grow corn for grain from those that could only produce silage, or winter wheat areas from spring wheat and flax regions, has traditionally been mapped as running east-west through the central Great Lakes region (Fig. 2A, B). Although the purpose of this paper is not to evaluate the efficacy of the choice of 8°C as a soil temperature regime boundary, hindsight has nonetheless been kind to this value/choice. Podzolization tends to become much stronger at soil temperatures below 8°, and the morphologies of Alfisols and sandy Entisols change markedly across these different temperature zones, with the region of most rapid change often coinciding with the M-F line (Smith, 1986; Schaetzl and Isard, 1996). Major vegetation boundaries in the central Great Lakes region also generally coincide with the M-F line, or lie within a few tens of kilometers of it (Elliott, 1953; Curtis, 1959; Comer et al., 1995; Medley and Harman, 1989; Schaetzl and Isard, 1991).

Early, published maps of the M-F line placed it within central lower Michigan and central Wisconsin (Fig. 2A and 2B). Our modeled data show that these early attempts at siting the line were reasonably accurate; our estimated M-F line, based on modeled data, is shown in Fig. 1, 2C and 3. The modeled location of the line generally parallels older estimates of its location, but turns noticeably north along the Lake Michigan lake effect snowbelt in lower Michigan. The mesic area within the Lake Michigan snowbelt is delineated today by MLRA 96 (Western Michigan and Northeastern Wisconsin Fruit Belt), which was traditionally viewed as an area of frigid soils (USDA-SCS, 1981), but is now being mapped by the NRCS using mesic soil series definitions, largely due to the efforts (and early products) of our research (Isard and Schaetzl, 1995). The M-F line currently being used by NRCS soil scientists in Michigan, that is, the eastern boundary of MLRA 96, is shown on Fig. 2B as a dotted line. The currently NRCS-accepted M-F line in lower Michigan continues eastward across the peninsula in generally the same location as the one established in 1981. In Wisconsin, there has been little or no research on the location of the M-F line; NRCS personnel there are using the M-F line that was established by the NRCS in 1993 (Fig. 2B).

It should be noted that our modeled M-F line is based on data from upland, forested soils; anecdotal evidence suggest that wetter sites are generally cooler, as are cultivated areas (Schaetzl and Tomczack, 2002). Because Soil Taxonomy (Soil Survey Staff, 1999) does not explicitly state that soil temperature measurements should be taken from a particular drainage class or land cover

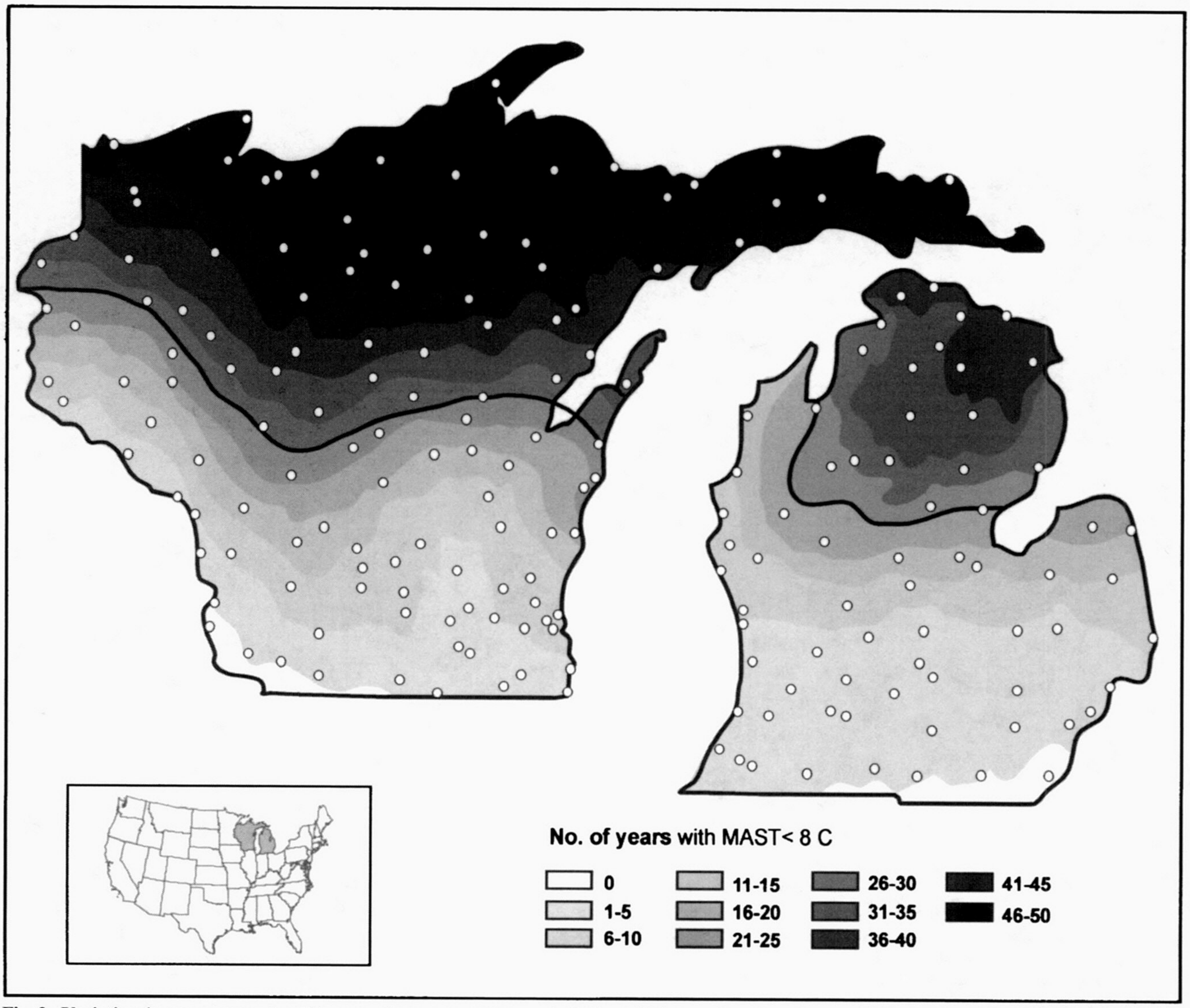

Fig. 3. Variation in mean annual soil temperatures across the central Great Lakes region, based on model output, using climatic data for the period 1951–2000; climatic data stations used in the analysis are indicated by small circles. The shaded choropleth map indicates the number of years in which the modeled mean annual soil temperature (MAST) was < 8°C for the period 1951–2000. The line shown is the best estimate of the mesic-frigid (M-F) boundary, based on 50% of the years being greater than 8°C.

type, we opted for upland sites that exist under forest cover, which was the dominant cover type at the time of European settlement and under which the soils of this region developed. Thus, any estimation of the M-F line in this region (or elsewhere) based on wetter soils or cultivated soils may be slightly different than our findings.

The modeled M-F line also follows topography to a certain extent, with cooler "highlands" being to the north and east of the line in both Michigan and Wisconsin (Fig. 2C). Our modeled M-F line is also substantially farther north in the western parts of Michigan and Wisconsin than earlier estimates had shown (Fig. 2). Based on our data, the soils in northern part of the Door Peninsula (WI) are frigid and the entire Leelanau Peninsula (MI) is mesic. Detailed inspection of data and sites along the east side of Grand Traverse Bay in Michigan suggests that the M-F line in that region may not end at the southern end of Grand Traverse Bay, as shown, but instead may run north, 10 to 20 km inland and east of the Bay, entering Lake Michigan just west of Charlevoix, MI (Fig. 2C). This interpretation is also supported by contemporary land use within this thin strip of land; orchards are successful here but fail to survive economically on frigid, higher elevation sites father inland. Because we are unsure of the exact location of the M-F boundary in this region, and because it is difficult to show this possible scenario at the scale of the maps used in this paper, we simply end the M-F line at the southern end of Grand Traverse Bay (Fig. 2C). Nonetheless, the general spatial agreement between earlier NRCS estimates of the M-F line location, and our modeled estimate, is striking. We think this bodes well for both our model and the NRCS'S soil temperature estimations nationwide.

Our modeled M-F line location, shown in Fig. 1, 2C and 3, represents the mean position of the 8°C soil isotherm for the 1951–2000 period. Variation about this line is an important attribute that cannot be ignored, and may figure centrally into studies of climate change

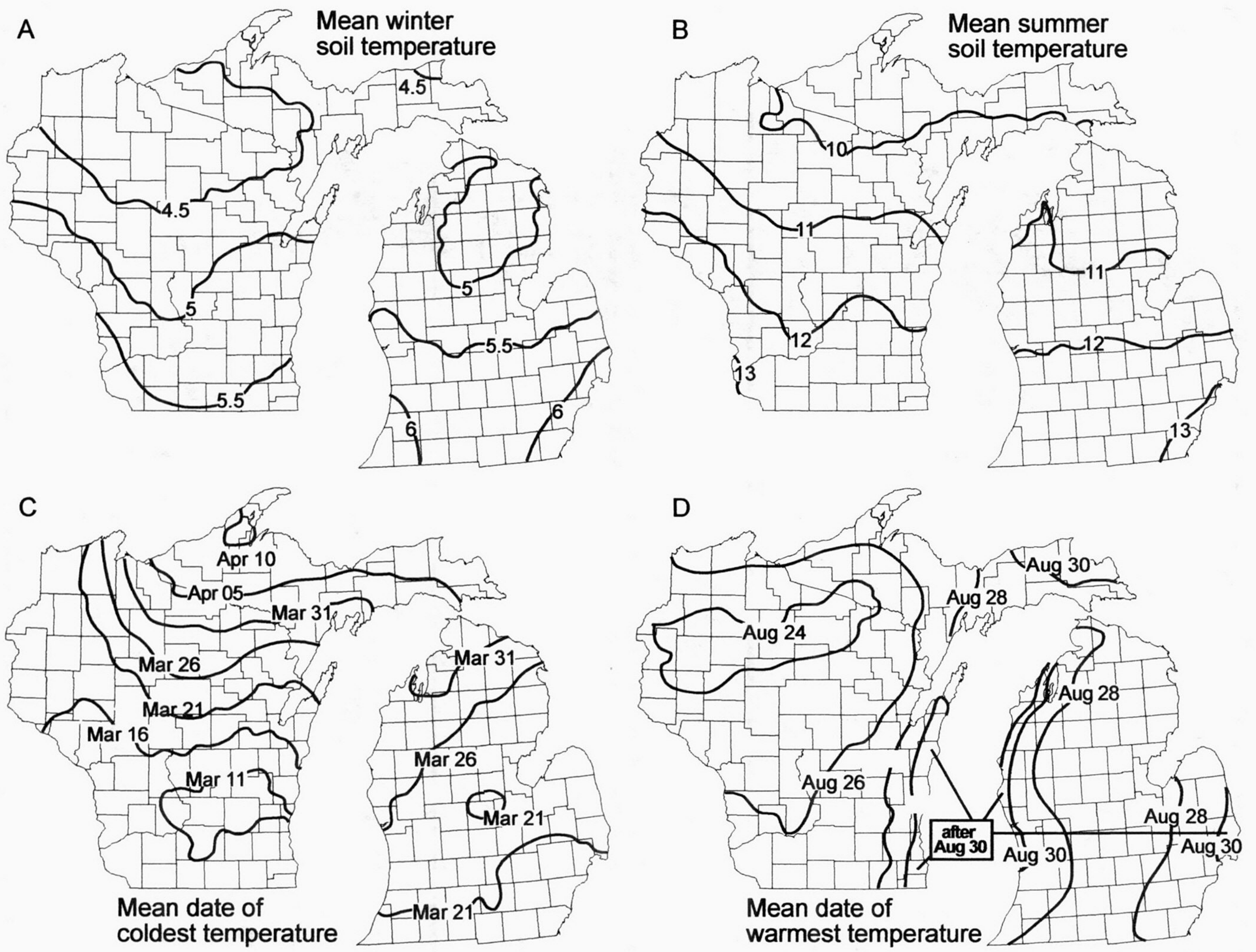

Fig. 4. Maps of variation in soil mean temperatures spatially and temporally, across the study area. A. Mean wintertime (November–April) soil temperatures (C) at 50-cm depth. B. Mean summertime (May-Oct) soil temperatures (C) at 50-cm depth. C. Mean date of coldest soil temperature at 50-cm depth. D. Mean date of warmest soil temperature at 50-cm depth.

across the region, which is projected to warm as greenhouse gases increase (Solomon and Bartlein, 1992; Iverson and Prasad, 1998). Figure 3 provides data on the interannual variation in MASTs across the region. Over a 50-yr timespan, soils in all but the southern tier of counties can expect to cool to below frigid levels (calculated on an annual basis) at least once. Likewise, the coldest parts of the region are the western upper peninsula (UP) of Michigan and the northern tier of counties in Wisconsin, which are frigid in >90% of the years. All sites in the lower peninsula of Michigan can expect to be mesic in at least 1 yr out of 10. Figure 3 also lends credence to the local belief that the area just north of Green Bay, known as the "Banana Belt of the UP" for its locally warmer climate, is indeed a warm outlier, as is the Garden Peninsula to its immediate east.

The variability map (Fig. 3) also provides data on soil temperature gradients. The strongest gradient in and near the M-F line occurs in western Wisconsin, whereas the soil temperature gradient is weakest in lower Michigan where it turns from an east-west to a north-south alignment. Indeed, the latter area is one of contention and low confidence with regard to soil temperature, for several reasons: (1) it exists at the junction between east-west trending gradients in air temperature and the north-south trending snowbelt (which is highly variable in position and snowpack depth from year to year), and (2) effects of topography on air temperature and snowfall totals in this region are complex, as this is an area of high, rugged interlobate moraines, such that along some reaches, the line cuts perpendicularly to topographic trends. Nonetheless, spatial gradients in soil temperature are influenced/confounded by more than topography and snowpacks, as gradients are also weak in the low-relief glacial lake plains of eastern lower Michigan as well.

Smith et al. (1964) were some of the first to observe that MAST is reasonably approximated by MAAT + 2°F, which is roughly equivalent to MAAT + 1°C. They pointed out instances in which this relationship breaks down: on steeply sloping terrain, in areas where the O horizon is thick, and in regions of heavy snowfall. For example, MAST under thick O horizons could be as cold or colder than MAAT (Smith et al., 1964). Isard and Schaetzl (1995) found that in southern Michigan, snow cover, especially in lake effect areas, insulates the

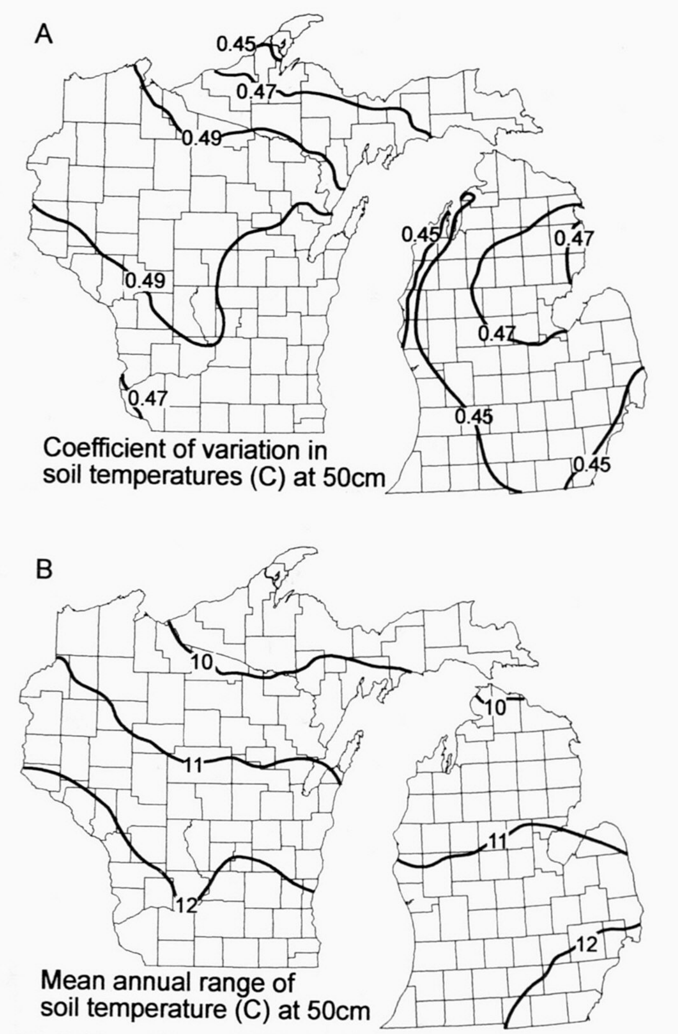

Fig. 5. Maps of the variability in soil temperatures across the study area. A. Coefficient of variation in soil temperatures at 50-cm depth. B. Mean annual range in soil temperatures at 50-cm depth.

soil, resulting in MAST values that exceed that of the [MAAT + 1°C] equation. Figure 2D shows the M-F line, as approximated by the [MAAT + 1°C] equation; our modeled line is shown in Fig. 2C. The [MAAT + 1°C] equation, the only option for remote areas that may be lacking soil temperature data but which do have climatological data, yields data that are somewhat different from our modeled data. The areas of greatest deviation are, surprisingly, not in the snowbelts, but on the eastern sides of Wisconsin and Michigan, where snowfall is minimal. In eastern lower Michigan, far from the Lake Michigan snowbelt, the [MAAT + 1°C] line is too far north, implying that this equation predicts that the soils are colder than they actually are. In both eastern and western Wisconsin, neither of which have lake effect snowbelts, the opposite situation occurs. It is difficult to draw conclusions from this analysis, except to say that [MAAT + 1°C] is a better approximator of the M-F line in this region than we had anticipated. Its utility for other, similar areas, is potentially very good, as long as it is recognized that the [MAAT + 1°C] line has a potential spatial error of up to 50 to 60 km in areas that accumulate thick snowpacks.

The maps in Fig. 4 and 5 illustrate some of the annual variation that exists in the soils of the study area. Mean winter temperatures are coldest in northern and northwestern Wisconsin, where air temperatures are often not moderated by having crossed the open waters of any of the Great Lakes. Snowbelt sites in Michigan's UP that are at the same latitude as non-snowbelt sites in northern Wisconsin are noticeably warmer, not only because the air is moderated as it crosses a much wider part of Lake Superior than it does in northern Wisconsin, but also because these areas are insulated by thick lake effect snowpacks. The effect of snowpacks on winter soil temperatures is evident in southern Michigan as well, where soil temperature isolines bend and become more meridional in and near the N-S trending lake-effect snowbelt that lies to the lee of Lake Michigan. A pocket of relative warmth also occurs in the western UP, centered on the lake effect snowbelt associated with Lake Superior. In summer (Fig. 4B), however, isolines are more zonal (E-W), following air temperature isotherms; there is no "snowbelt effect."

Because snowpacks persist in the spring, they keep the soils in those regions cold for longer periods of time than occurs in non-snowbelt areas by (i) insulating the soils from warm air that occasionally advects into the region in spring and (ii) periodic additions of cold snowmelt water (Fig. 4C). These two factors, coupled with the cold air associated with Lake Superior, lead to persistence of cold soil temperatures long into spring, in areas of the northern UP. A similar pattern occurs in southern Michigan, and again is associated with the Lake Michigan snowbelt; warm-up in spring is delayed and the date of coldest soil temperature occurs later in spring (Fig. 4C).

The spatial pattern of coldness of springtime soil temperatures in snowbelt areas has implications for pedogenesis. Schaetzl and Isard (1991, 1996) indicated that podzolization was enhanced by slow, continuous snowmelt infiltration, for a number of reasons: (1) it percolates through "fresh" litter (from the previous fall) that is able to contribute large amounts of organic acids, which aid in podzolization (Buurman and van Reeuwijk, 1984; Vance et al., 1985; Krzyszowska et al., 1996; van Hees et al., 2000), (2) evapotransipration demands are low and soils are often already wet at the onset of snowmelt, allowing wetting fronts to penetrate and translocate materials deeper, and (3) cold soil and soil water temperatures reduce microbial activity, which can act to break down any organo-metallic complexes that are percolating in solution, thereby stopping translocation (Lundström et al., 1995). Data on Fe and Al contents of soil solutions confirm the efficacy of podzolization during snowmelt (Schaetzl, 1990). It is, therefore, no coincidence that cold and persistently cold springtime soil temperatures coincide spatially with areas of intense podzolization in the Great Lakes region (Schaetzl and Isard, 1991, 1996).

Spatial patterns of summertime soil temperatures, as typified by the mean date of warmest soil temperature (Fig. 4D), are readily explainable. Soils warm most quickly and attain their warmest conditions earlier in the year at locations that are more "continental," that is, farthest from the moderating effects of the Great

Lakes. Locations near the lakes reach their warmest temperatures as much as a week later than sites inland.

Sites far from the Great Lakes also show more extremes of soil temperature variation, as indicated by coefficient of variation (CV) (Fig. 5A). Peninsulas and near-shore locations have lower CVs, indicative of the moderating effects of the lakes and the snowpacks associated with them. Annual range of temperature is less affected by the lakes, showing a more E-W pattern, with soils in the south having larger annual ranges of temperature (Fig. 5B).

CONCLUSIONS

Air temperature data, for 1951–2000, from 218 stations in the central Great Lakes region were input into a model that outputs soil temperatures at 50-cm depths. Verification of the model's accuracy was based on comparisons to data from well-drained, forested soils throughout northern lower Michigan; the mean biased error of the model was $<0.1°C$. Maps of the modeled M-F line were developed, and show good spatial agreement with earlier estimates of the line by the NRCS, suggesting that the time-honored equation whereby MAST = MAAT + 1°C is reasonably accurate in this region, despite deep snowpacks and year-round lake effects. When compared with sites inland, locations near the Great Lakes tend to (i) be noticeably warmer in winter, (ii) stay colder longer into spring and warmer longer into fall, and (iii) experience less intra-annual variation in soil temperature, as expressed by coefficients of variation.

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