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Regional-scale relationships between climate and strength of podzolization in the Great Lakes Region, North America

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Abstract

Along a 300 km transect in northern Wisconsin and the upper peninsula of Michigan, the areal coverage of Spodosols and spodic horizon development increases markedly from south to north. This study elucidated those aspects of climate that promote podzolization in this region, through an examination of the geographic correspondences between Spodosol development and soil climate. Climatic data (1951–1991) from 21 sites along this transect were processed by a hydrologic model developed to output data on (1) soil temperatures for 0.05 and 0.5 m depths for 20 minute intervals, including data on soil freezing, (2) snowpack thicknesses under forest cover, and (3) daily water fluxes, runoff, and soil water contents at several depths.

Spodosols dominate the landscape in areas where soil frost and freeze–thaw activity are minimal and where soil temperatures rarely exceed 16–17°C. Podzolization is strongest where snowpacks are thickest — an association that holds at both regional and meso scales. Thick snowpacks inhibit soil frost and allow large fluxes of snowmelt water to infiltrate into already moist profiles. This type of flux (slow, steady, cold water) may be particularly effective in the podzolization process. In the southern part of the transect, where Spodosols are rare, snowmelt fluxes are 1/3 as large as in the northern ''snowbelt'' areas. The southern areas also have a small autumn infiltration peak that usually reaches to ≈ 0.3 m depth; this flux is absent in areas of strong podzolization. Mean soil water contents are low and fluxes of water into the soil are small along the entire transect during summer, underscoring the belief that the bulk of pedogenesis (i.e., translocation), in Spodosols in the study area, occurs during snowmelt.

Keywords: soil genesis; snowpack; snowmelt; Spodosols

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1. Introduction

Podzolization as a pedogenic process has been long studied (Muir, 1961, Petersen, 1976). Podzols and Spodosols (Canadian Soil Survey Committee, 1978; Soil Survey Staff, 1992) are found in loamy or coarser materials, and form in humid climates (Steila and Pond, 1989). These soils generally have an eluvial zone which has lost organic matter and Al, and possibly Fe. The illuvial zone below is typically reddish or brown-black, due to accumulations of organic matter and/or Fe. They are widespread in subpolar and alpine climatic regimes, and in some areas extend into interiors of the mid-latitude continents (Steila and Pond, 1989; Padley et al., 1985).

Traditionally and in theory, the following sequence of processes are postulated as major components of podzolization: acidification and chemical weathering, release of Fe and Al cations into the soil solution, mobilization and translocation of sesquioxide–organic chelate complexes, and deposition in illuvial (B) horizons (Mokma and Buurman, 1982; McKeague et al., 1983; Buurman and Van Reeuwijk, 1984; Petersen, 1984; Ugolini and Dahlgren, 1987). Another theory purports that much of the translocation of the Fe and Al is accomplished as inorganic Fe oxide–silica–alumina–water sols, later to be precipitated as allophane and imogolite. Later, organic matter may precipitate onto these clay minerals and short range ordered minerals (Anderson et al., 1982; Farmer, 1982; Childs et al., 1983). Ugolini et al. (1990) have proposed a theory that incorporates elements of the above two.

Regardless of which scenario is more applicable, a vehicle by which these translocations occur must be operative or spodic (Podzolic) morphologies cannot form. This vehicle or mechanism is downward-moving (infiltrating) water. Water provides the energy for many horizonation processes (cf. Runge, 1973). Thus, Spodosol profiles form only under leaching conditions, in humid or perhumid climates (McKeague et al., 1983), where an excess of precipitation over evapotranspiration (ET) exists for at least some period of the year. Deep leaching is enhanced in coarse-textured parent materials. These coarse-textured materials usually have an abundance of quartz and low amounts of iron and weatherable minerals. Thus, such materials are more rapidly weathered and stripped of ions and other coatings, on a unit volume basis, than are Fe- and clay-rich sediments (Duchaufour and Souchier, 1978), enabling a bleached eluvial horizon, rich in quartz, to form comparatively rapidly. For this reason, podzolization is favored in sandy and gravelly parent materials, and is slow to occur in loamy or finer-textured parent materials.

Vegetation impacts podzolization because species that produce abundant, acidic litter will accelerate the weathering and chelation processes more than those whose litter is not especially acidic (Ugolini et al., 1990). Examples of the former include heath, Rhododendron, and most coniferous species, especially hemlock (*Tsuga* spp.), fir (*Abies* spp.), spruce (*Picea* spp.) and pine (*Pinus* spp.).

Little previous research on podzolization has focussed on the impact of climate, since vegetation, parent material and other factors usually either dominate the climatic factor, or are more easily isolated in a "soil sequence" study (Sevink, 1991). Thus, pattern–process studies of podzolization have traditionally focussed on factors other than climate (with few exceptions, cf. Jauhiainen, 1973). For example, podzolization has been studied in relation to factors of vegetation (Evans, 1980; Crampton, 1982), parent material (DeConinck and Laruelle, 1960; Tonkonogov, 1971), time (Franzmeier et al., 1963; Barrett and Schaetzl, 1992), and relief (Wang et al., 1991). Less often, however, has climate been the focus of podzolization studies, perhaps because of the changing nature of climate over time. Nonetheless, even though climate *does* change, so does vegetation, and the impact of changing vegetation may be more sudden and cause more rapid morphological changes than that of climate (Hole, 1975; Sohet et al., 1988).

The purpose of this research is to identify climatic elements, both atmospheric as well as pedologic (Jenny, 1941), that accentuate the podzolization process in the Great Lakes region. We achieve this through a regional-scale, geographic examination of soils and climate in northern Michigan and northern Wisconsin, across which the abundance of Spodosols and the strength of spodic horizon expression changes markedly and systematically. In previous work in southern Michigan (Schaetzl and Isard, 1991; Isard and Schaetzl, 1995) we noted a strong spatial coincidence between areas of strongly-developed Spodosols and climatic factors including: (1) high amounts of infiltration during snowmelt (March and April), (2) relatively dry soil conditions in autumn, and (3) low potential for soil frost. Herein we expand on this work by extending our model to track downward movement of soil water, and by reporting modelled data on soil freezing.

2. Study area and soil geography

The study area is located in northern Wisconsin and the western part of Michigan's upper peninsula (Fig. 1). The pronounced decrease in Spodosol abundance from north to south across the region is depicted by USDA–SCS STATSGO (State Soils Geographic Data Base) data on the percentage of polygons covered with Spodosols (Fig. 2A). A similar, though less apparent, trend occurs in strength of Spodosol development (Fig. 2B).

A close inspection of the two maps in Fig. 2 reveals that the regional distribution of Typic Spodosols is less homogeneous than that of Spodosol frequency. This is probably due to variation in soil-forming factors other than climate, which can change markedly across short distances. These sorts of short-distance changes in morphology, however, are not addressed in our study of *regional* trends in Spodosol development.

Vegetation in the region consists of mixed, coniferous-deciduous forests (Graham, 1941; Curtis, 1959; Goff, 1967; Webb et al., 1983; Barrett et al., 1995). Well-drained, sandy soils generally support a higher percentage of coniferous species than do well-drained, loamy soils (Curtis, 1959; Whitney, 1986; Barrett et al., 1995). Although there may exist a slight south to north trend of increasing coniferous component to the forests (with a corresponding increase in acidic litter production), it is not likely that the latitudinal gradient in vegetation alone is responsible for the dramatic pedologic changes along the same transect.

Most soils have formed in deposits of glacial or fluvioglacial origin. The looping terminal moraine of the Wisconsin Valley, Langlade, and Green Bay glacial lobes marks the southern limit of Woodfordian ice (Fig. 2A). This ice was probably at its maximum extent approximately 13,000 yr BP (Maher, 1981; Mickelson et al., 1983). South and

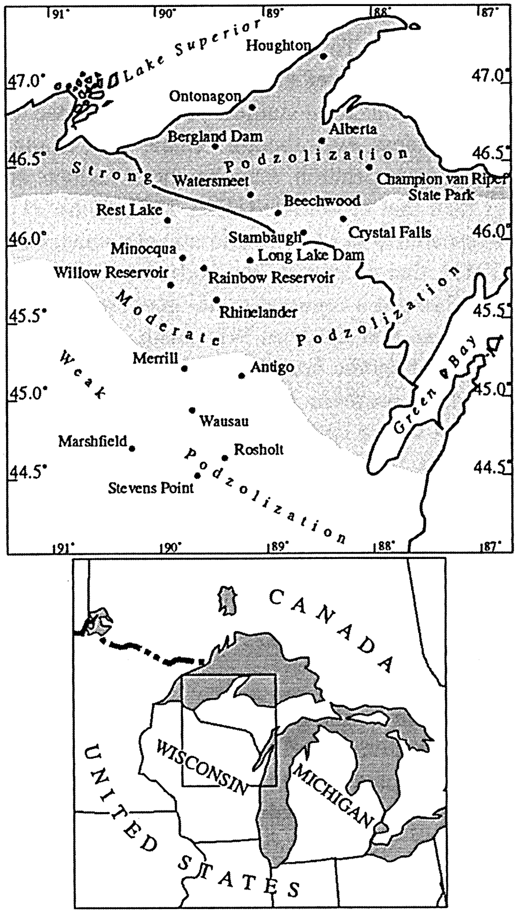

Fig. 1. Map of the study area. National Weather Service stations utilized in the analysis are indicated. Strength of development of Spodosols (indicated by shading), generalized from data in Fig. 2.

west of this moraine are glacial deposits of Pre- and Early-Wisconsin (> 80,000 yr BP?) age (Mickelson et al., 1983, 1984), as well as some younger, Woodfordian-aged outwash deposits. Many of the older sediments are fine-textured (loams and clay loams) and may contain a thin loess cap. North and east of the Woodfordian moraine, gravelly, sandy and loamy glacial deposits dominate the landscape (Hole, 1976; Mickelson et al., 1983). The northernmost parts of the study area, in northern Michigan, were deglaciated by ≈ 9800 BP (Futyma, 1981).

In the western two-thirds of the study area, the drift was derived primarily from acid, Precambrian rocks. To the east, a significant component of limestone and dolomite was mixed into the tills, rendering them initially calcareous (Mickelson et al., 1983). Nonetheless, because most of the sediments in the area are coarse-textured outwashes and tills, podzolization has been active. Whereas sediment texture and carbonate content

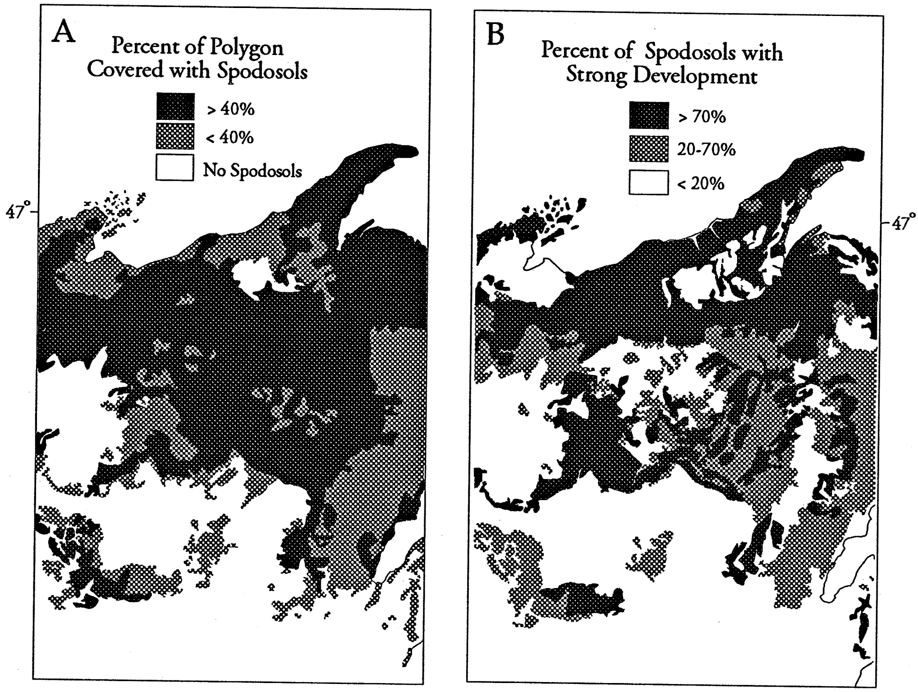

Fig. 2. (A) Location of Spodosols in study area, as indicated by the percentage of soil polygons that are Spodosols (STATSGO, USDA–SCS, Lincoln, NE, 1994). (B) Location of Spodosols with strong (Typic) development, as indicated by the percentage of Spodosols that are either Typic or Alfic Haplorthods, or Typic or Alfic Fragiorthods.

has affected the rate of podzolization at some sites, at a regional-scale, the north to south trend in areal coverage of Spodosols (more Spodosols in the north) is very prominent and clearly overwhelms most small-scale variations in parent material. This point is supported by work on the geography of Podzols in northern Russia, in which climate and parent material were both shown to interact to produce distinct geographic patterns (Tonkonogov, 1971). The importance of the influence of climate (specifically, *bioclimate*) vs. parent material on these patterns is, however, affected by the *scale* at which the patterns are studied. At our *regional* scale of analysis, in which climatic gradients are evident, we theorized that this soil-forming factor (Jenny, 1941) would overwhelm parent material or vegetation in producing broad geographic variations in Spodosol distribution. At a regional scale of analysis, therefore, all other soil-forming factors can be considered as being relatively constant.

We assume that, although the specific *values* of important climatic factors have changed during the Holocene (e.g., Webb and Bryson, 1972), the *patterns* of recent climate are similar to those occurring during the Holocene. For example, many of the patterns of climatic factors are influenced by Lake Superior, and we assume that the lake's effect on climate at the regional-scale has not changed since the last retreat of the ice.

Poorly-drained and wetter Spodosols are not included in this analysis because th presence of a water table near the surface greatly modifies translocations of soi components.

3. Methods

3.1. Soil water and temperature model

Vertical profiles of water and temperature were calculated for coarse-texture (coarse-loamy and coarser), well-drained, forested soils using a modified form of th soil water and temperature computer algorithm developed previously (Schaetzl an Isard, 1991, Isard and Schaetzl, 1995). The model uses a Newhall-based, water budge component (Van Wambeke et al., 1986), combined with the Soil Conservation Service' (SCS) snowmelt model (USDA–SCS, 1971) and a one-dimensional heat conductio equation (Carslaw and Jaeger, 1959). The model is formulated with twenty soil layers each 5 cm thick, five additional soil layers that increase in thickness with increasin depth, one litter layer, and up to ten snowpack layers (Table 1). The model assume homogeneous, piston flow of infiltrating water into soils, and although this has bee shown not to be the case even in relatively homogeneous sands (e.g., Price and Bauer

Table 1
Parameters for litter and soil layers used in the hydrologic model

Compartment evaluation order					Available water (mm)	Location and thickness of layer
–	–	–	–	–	2.0	0–5 cm above soil surface (litter; O horizon)
–	–	–	–	–	10.5	0–5 cm depth (mineral soil)
11	7	4	2	1	1.2	5–10 cm depth
16	12	8	5	3	1.2	10–15 cm depth
21	17	13	9	6	1.2	15–20 cm depth
26	22	18	14	10	1.2	20–25 cm depth
31	27	23	19	15	1.2	25–30 cm depth
36	32	28	24	20	1.2	30–35 cm depth
41	37	33	29	25	1.2	35–40 cm depth
46	42	38	34	30	1.2	40–45 cm depth
51	47	43	39	35	1.2	45–50 cm depth
56	52	48	44	40	0.7	50–55 cm depth
61	57	53	49	45	0.7	55–60 cm depth
66	62	58	54	50	0.7	60–65 cm depth
71	67	63	59	55	0.7	65–70 cm depth
76	72	68	64	60	0.7	70–75 cm depth
81	77	73	69	65	0.7	75–80 cm depth
86	82	78	74	70	0.7	80–85 cm depth
90	87	83	79	75	0.7	85–90 cm depth
93	91	88	84	80	0.7	90–95 cm depth
95	94	92	89	85	0.7	95–100 cm depth

Layers below 100 cm were not used in water flux calculations.

Table 2
Parameters values for sensitivity analysis

Model parameter [a]	Normal model runs	Sensitivity analysis	
		level A	level B
Litter thickness (mm)	50	25	100
Forest stemflow plus throughfall (mm)			
for solid precipitation [b]	$0.7 \cdot P$	P	$0.5 \cdot P$
for liquid precipitation [c]	$-2.33+0.859 \cdot P$	$-0.04+0.81 \cdot P$	$-1.70+0.63 \cdot P$
Water storage capacity (mm): [d]			
@ 5–50 cm	1.2	0.7	5.0
@ 50–100 cm	0.7	0.7	1.2
Snowmelt (mm) [e]	$2.2176 \cdot T_{avg}$	$1.828 \cdot (T_{max} - 5.556)$	$0.914 \cdot (T_{max} - 5.556)$
Soil thermal properties: [f]			
thermal conductivity ($cal\ cm^{-1}\ s^{-1}\ K^{-1}$)	$0.0054/[1+6.714 \cdot \exp(-4.41 \cdot SMR)]$	0.0007	0.0052
heat capacity ($cal\ cm^{-3}\ K^{-1}$)	$0.3+0.42 \cdot SMR$	0.3	0.7

[a] P is liquid precipitation, T_{avg} is the average daily temperature (°C), T_{max} is maximum daily temperature (°C), and SMR is the ratio of soil water content, as calculated by the model, to water storage capacity in the soil layer (Table 1).
[b] Wood (1937); Hansen (1969); Satterlund and Haupt (1967); Schaetzl and Isard (1991).
[c] Voigt (1960); Helvey (1967); Rogerson and Byrnes (1968); Schaetzl and Isard (1991).
[d] Soil Conservation Service (1980).
[e] Garstka (1944).
[f] Lowrey and Lowrey (1989).

1984, Kung, 1990), for this level and scale of analysis it nonetheless provides a representation of average fluxes of soil water over periods from days to years.

National Weather Service (NWS) daily maximum and minimum air temperatures and precipitation data were used as inputs to the computer model. 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 (Table 2). The water is stored in a snowpack when throughfall and air temperature conditions allow. Snowmelt is calculated as a function of air temperature (Table 2) 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 are specified in Table 1, and are based on data collected in the field (Schaetzl and Isard, 1990). The model, modified from Van Wambeke et al. (1986), simulates the progression of a wetting front into the soil. For this computation, each soil layer is subdivided into five compartments and allotted one-fifth of the water available in the layer. The distance that the wetting front moves downward depends on the amount of water needed to bring storage in the litter and soil layers above to field capacity. Runoff or lateral throughflow out of the soil system may occur when the temperature in the litter layer is $< 0°C$ or when a soil layer is frozen and water storage is at capacity in the unfrozen layers above.

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, additional water is removed from lower layers following the procedure suggested by Van Wambeke et al. (1986). Water is withdrawn from the compartments in the order specified in Table 1, assuming a linear relationship between the ratio of water removal to potential evapotranspiration and available water (Baier and Robertson, 1966).

Temperature in the lowest soil layer (7–15 m depth) is held at 2°C above the mean annual air temperature, as suggested by Geiger (1965) and Smith et al. (1964). Air temperature at the litter or snow surface is 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 2 pm Central Standard Time, and daily minimums at dawn. Thermal properties for the 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 content (Table 2) as suggested by Lowrey and Lowrey (1989). A finite difference formulation is used to calculate the temperature profile within the mineral soil at 20 minute intervals. In many respects, our model is similar to the one used by Brooks et al. (1995) to model soil climate.

The algorithm was evaluated by comparing its output to approximately 500 soil temperature observations collected at multiple depths between, 1990 and 1993 from 10 sites in Michigan (Isard and Schaetzl, 1995). The root mean square errors and mean bias of the errors in the soil temperature simulations ranged from 1.1 to 2.5°C and from $+0.6$ to -1.6°C, respectively, depending upon depth and season.

3.2. Climatic data

Soil water fluxes and temperatures were simulated by the model, using climatic data from 21 NWS stations in northern Wisconsin and upper Michigan for the period between 1951 and 1992. The stations range in latitude from 44.50° (Stevens Point, WI) to 47.17°N (Houghton, MI), comprising a broad north–south transect of approximately 300 km (Fig. 1). Because the 1951–1992 data for most stations contained occasional missing values, we developed a "buddy" system for estimating missing data. Each of the 21 primary stations was assigned three buddy stations. Buddies 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 a surrogate from the second, and if necessary, the third buddy station.

3.3. Sensitivity analysis

An analysis was performed to evaluate the sensitivity of infiltration and soil temperature calculations to changes in key model parameters for the soil (litter depth, water

Table 3
Summary of sensitivity analysis results

Output variables	Row No.	Model input parameters				
		Litter thickness	Throughfall equations	Water storage capacity	Snowmelt equations	Soil thermal properties
Monthly soil water flux @ 0 cm	1	moderate	moderate/strong	little	moderate/strong	none
	2	growing season	year round	year round	winter only	
	3	similar response among stations	similar response among stations	similar response among stations	divergent response among stations	
Monthly soil water flux @ 35 cm	1	very little	moderate	none	strong	none
	2	growing season	year round		winter only	
	3	similar response among stations among stations	similar response among stations		divergent response	
No. of freeze–thaw cycles/mo @ 10 cm	1	none	strong	very little	strong	moderate
	2	Sept.–May	October and April		Sept.–May	Sept.–April
	3	divergent response among stations			divergent response among stations	divergent response among stations

Row numbers have the following meaning: 1. This row details the effects of varying the model's input parameters on the output variable values. See Table 2 for input values used in the sensitivity analysis. 2. For output variables which changed in response to input parameter variations, this row details the time of year for which that change was most pronounced. 3. This row reveals whether the change in the output variable values were similar or dissimilar (divergent) among northern vs. southern stations.

storage [field] capacity, and thermal properties) and forest (coefficients in canopy throughfall and snowmelt equations). Model performance was evaluated using a range of values reported in the literature for each of the model parameters (Table 2) for three stations within the study area: Marshfield, WI (44.65°), Minocqua, WI (45.88°), and Ontonagon, MI (46.83°N) (Fig. 1).

A brief summary of the sensitivity analyses is presented in Table 3, including an indication of how sensitive the output variables are to changes in each parameter, the time of year for which the model response is most pronounced, and whether the three stations at different positions along the transect displayed similar or divergent responses to the changes in parameter values. The analysis revealed that the downward flux of soil water is affected most by varying the input parameters associated with snowmelt, and to a lesser extent, throughfall (Table 3). Although there was a divergent response in infiltration values for the three stations to changes in the snowmelt equation, the general regional-scale patterns of infiltration remained nearly the same. Changing the soil thermal properties in the hydrologic model (thermal conductivity and heat capacity) produced divergent, though not large, differences in the predicted soil temperatures (Table 3). Snowmelt and throughfall equations affected freeze–thaw cycle frequencies from September through April, because both directly impact snowpack thicknesses. The model is more likely to under-estimate than overestimate snowpack thicknesses; we compared modelled snowpack thicknesses with snowpack depth measurements in Michigan forests (Isard and Schaetzl, 1993) and are reasonably satisfied with the comparisons.

3.4. Graphical presentation

Precipitation data and values of soil climate factors calculated by the model were plotted, with month on the *x*-axis and latitude on the *y*-axis using DeltaGraph® software. Smoothing of isolines was performed by hand to produce the final figures.

4. Results and discussion

4.1. The distribution of Spodosols

Fig. 2A shows the general distribution of Spodosols across the study area. The map shows that most Spodosols are found in the northern part of the study area, and that most of the Spodosols with strong development (Typic subgroups of Spodosols) are found in the northern third. Few Spodosols exist in the following areas, primarily due to the effects of parent material limitations: (1) south of the Woodfordian border in Wisconsin and in western north-central Wisconsin, where silty sediments (loess) cover much of the landscape, (2) in the extreme eastern and southeastern parts of the study area, where the tills are calcareous and, in places, soils are shallow to limestone bedrock, (3) in and near Lake Superior and in a small area in the west-central upper peninsula, where clayey lacustrine sediments dominate the landscape (Hole, 1976; Michigan State University Agricultural Experiment Station, 1981; Schwenner, 1989; Madison and Gundlach, 1993). In other areas, some Spodosols are usually found (Fig. 2). The

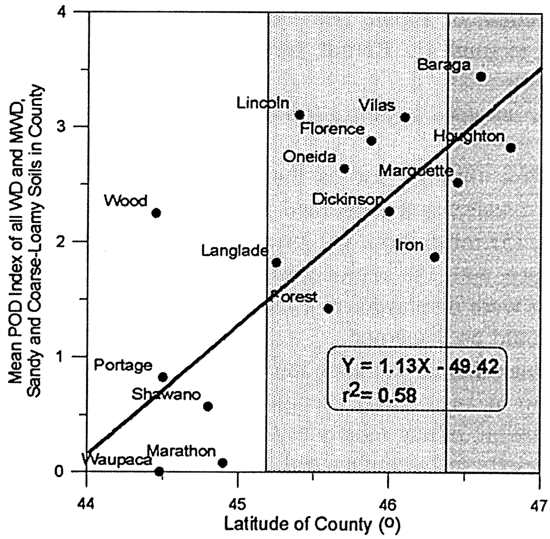

Fig. 3. Relationship between mean POD Index values and latitude of the center of the county, for most counties in the study area. Only moderately well-drained (or drier) soils in sandy, sandy-skeletal and coarse-loamy families were included in the analysis. Pedons that lacked an E horizon due to plowing were not considered, as per Schaetzl and Mokma (1988). In this and all subsequent figures, white-light gray-dark gray colors correspond to the areas shaded in Figure 1, and thus represent areas of strong, moderate, and weak podzolization.

strongest area of podzolization is north of about 46°N latitude. Most well-drained soils south of 45°N show little or no evidence of translocation of humus and/or sesquioxides, irrespective of parent material texture or pH.

As an additional measure of podzolization, we examined all soil profile descriptions from the 16 county-level soil surveys that have been completed for the study area. For each county, we calculated the POD Index on all moderately well-drained or drier pedons described therein, which were coarse-loamy or coarser in texture. The POD Index is a numerical estimate of strength of Podzolic soil development, and can be calculated from morphologic data by comparing E and Bhs (or Bs) horizon colors (Schaetzl and Mokma, 1988). Soils with virtually no evidence of podzolic or spodic development typically have POD Indices near zero, whereas PODs for strongly developed Orthods and Humods may range from 6 to >20. For example, Schaetzl (1992) described a Haplohumod in the central upper peninsula of Michigan with a POD Index of 8.5. The POD Indices were averaged and the resulting mean POD values regressed using latitude as the independent variable (Fig. 3). Fig. 3 thus provides a measure of Podzolic (more correctly, *Spodosolic*) soil morphology for the area, indicating that the mean POD Index ranges from near zero for counties in central Wisconsin to almost four near the southern shore of Lake Superior.

The trend of increasing strength of podzolization from south to north, shown in Figs. 2 and 3, is manifest not only across the region but also at the county scale. In Houghton

County, MI, the strength of spodic (Bhs) horizon development has been observed to decrease from north to south (L. Berndt, pers. commun., 1992).

4.2. Climatic factors involved in podzolization: water

Because water flow through soil is the driving force behind most translocations (Runge, 1973), we created and examined maps of: (1) total precipitation, (2) large precipitation events, (3) infiltration, (4) large infiltration events, and (5) soil moisture, for spatial congruence with the regional-scale maps of podzolization presence in the study area. The data were displayed on a monthly basis so that a temporal signature, if any were present, could be discerned.

Precipitation is maximal in summer throughout the study area, with most stations receiving > 100 mm mean monthly precipitation during the June–August period (Fig. 4A). Winter precipitation, falling mostly as snow, is maximal in the strong podzolization region (north), especially in December and January. Nonetheless, there does not appear to be a large difference in amount and timing of precipitation among the three podzolization zones.

In previous studies (Schaetzl and Isard, 1991; Isard and Schaetzl, 1995), we found that large inputs of water to the soil, although usually infrequent, appear to trigger and promote podzolization more effectively than numerous small inputs. Thus, we first examined precipitation records for precipitation events larger than 20 and 30 mm. Large precipitation events are more numerous in the southern two-thirds of the study area, probably due to intense rain from convectional summer thunderstorms, and from prolonged frontal rain in other seasons, both of which are more common in the south. Much of the atmospheric moisture that falls as precipitation in central United States has as its origin the Gulf of Mexico, and consequently, proximity to the Gulf is an important control over the latitudinal gradient of heavy rainfall events across the study area. The *amount* of precipitation derived from these large events, both the > 20 mm and > 30 mm events (Fig. 5B shows the pattern for the > 20 mm events only), however, is maximal in the moderate podzolization zone. Spatial patterns of the frequency of large precipitation events (also not shown but having a similar pattern to the data portrayed in Fig. 5B), as well as the magnitude of these events (Fig. 5B), while potentially important for supplying summer ET demand, do not correspond to regional-scale variations in the frequency of occurrence of Spodosols, and alone do not appear to be driving the podzolization process.

In our model approximately 60–70% of gross precipitation is allowed to pass though the forest canopy and is input to the soil surface as throughfall and stemflow. These values vary little from north to south. Canopy interception losses of 30–40%, as modelled here, compare favorably to measurements from similar forests elsewhere (Willis et al., 1975; Freedman and Prager, 1986). The ratio of water entering the mineral soil to gross precipitation can be used to illustrate the large losses of water when canopy interception is combined with evaporation from the litter. In mid-summer, less than 25% of gross precipitation enters the mineral soil throughout the entire study area, whereas during snowmelt, values for infiltration near Lake Superior exceed 100% of gross precipitation because of ''stored'' water in the snowpack.

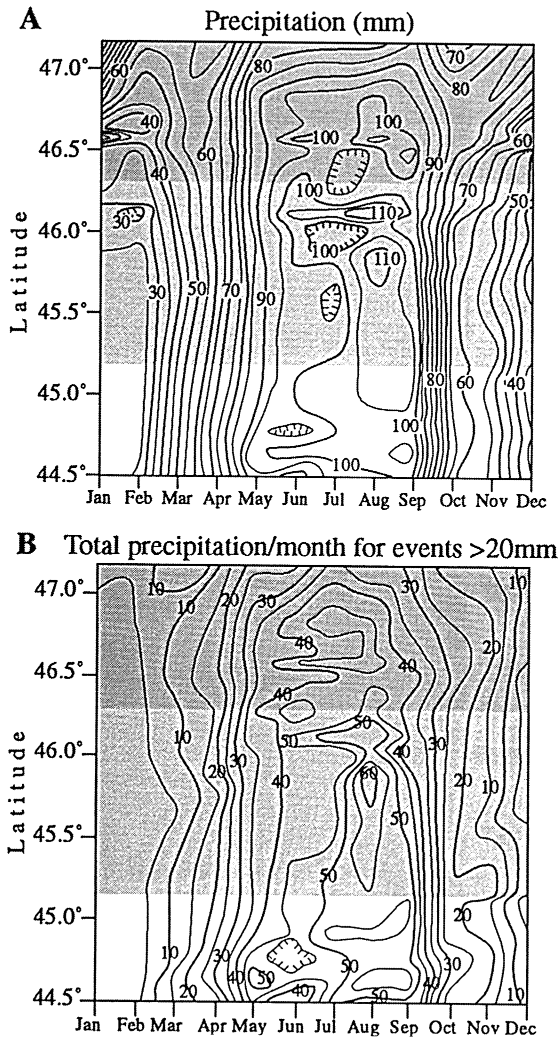

Fig. 4. (A) Mean monthly precipitation in study area from 1951 to 1992. (B) Total precipitation per month for events larger than 20 mm in study area, averaged over the 1951 to 1992 period of record.

Infiltration, at all depths and for all regions, is maximal during March and April (Fig. 5). Even though summer is a period of large precipitation events (Fig. 4B), much of this precipitation is caught on the forest canopy and in the litter layer and lost to evaporation prior to entering the soil (cf. Helvey and Patric, 1965; Mahendrappa and Kingston, 1982). Soils in the strong podzolization zone experience a large flux of water during snowmelt, due to deep snowpacks (Fig. 6) and rain-on-snow events. This trend has been noted previously for northern (Schaetzl and Isard, 1990) as well as for southern Michigan (Isard and Schaetzl, 1995). Soils in the zone of strong podzolization receive, at all depths, 3 to 4 times more infiltration during March and April than do soils in the

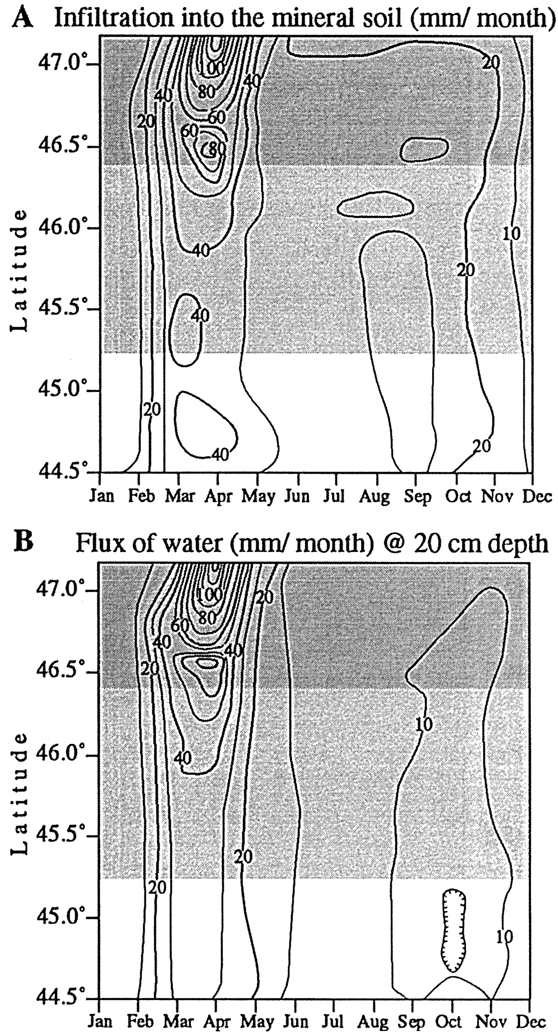

Fig. 5. Total monthly downward flux of water (modelled) through the: (A) mineral soil surface (0 m), and (B) 0.2 m depth, in the study area, averaged over the 1951 to 1992 period of record. The spatial patterns of the water flux varied little below 0.2 m.

southern zone of weak podzolization (Fig. 5), because of the high amounts of winter precipitation stored in snowpacks and released by spring melt (Fig. 6). However, even in the south, a small snowmelt infiltration peak is present (30–40 mm mo^{-1}). This southern peak is comparable to an early autumn infiltration peak (30 mm mo^{-1} in September). The early autumn peak is not observed in the north, and diminishes with depth in the south, such that it is no longer noticeable at depths below 0.5 m. In southern Michigan, Schaetzl and Isard (1991) also noted that areas of strong podzolization

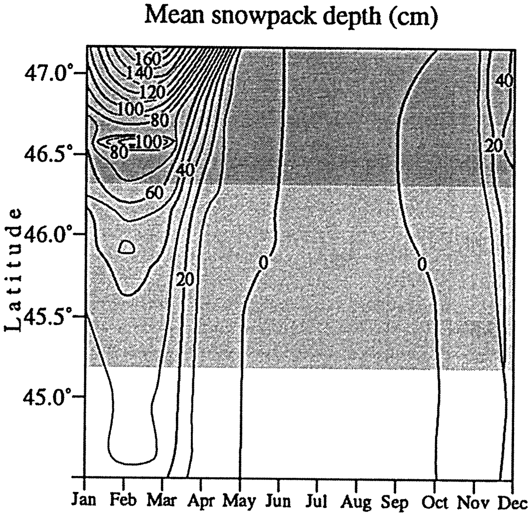

Fig. 6. Mean monthly snowpack depth (modelled) in study area for the 1951 to 1992 period of record.

experienced substantially more snowmelt infiltration and had drier autumns than did nearby areas with less developed soils.

As mentioned previously, the magnitude of infiltration events has been found to impact intensity of spodic expression in a given area, to a greater extent than the overall quantity of infiltrating soil water (Schaetzl and Isard, 1990). The average number of daily infiltration events > 13 mm per year show the frequency with which the very largest infiltration events occur, and are maximal during snowmelt in the strong podzolization region (Fig. 7). Here, during the 42 years of record, infiltration events of ≥ 13 mm occurred on average at least once per month during March through May, and as often as 8–10 times per month. It is likely that many of these large infiltration events occurred when abnormally warm air masses advected over a snow-covered surface, triggering large snowmelt infiltration events. In weak podzolization areas to the south, such large events occur, at most, once every two to five years, in large part because the ground is seldom snow-covered. Nonetheless, even here such large events are most likely to be a result of snowmelt rather than thunderstorms or prolonged rains.

Soil water as a percent of saturation is shown for the 0.05–0.1 m layer in Fig. 8. Maps of soil water as a percent of saturation for other soil layers were similar and have not been reproduced here. The rapid loss of soil water in April and May is due to transpiration after tree leaf-out. A mid-summer dry period is apparent across the entire study area, suggesting diminished weathering and translocation (i.e., podzolization) during this time. There is little latitudinal variation in soil moisture along the transect except in the extreme north, near Lake Superior, where the soil is relatively dry during winter.

Most areas in northern Minnesota, which is farther north than the area studied here, have weakly developed Spodosols (United States Geological Survey, 1985). This could be attributed, in the first instance, to finer parent materials and wetter sites. However,

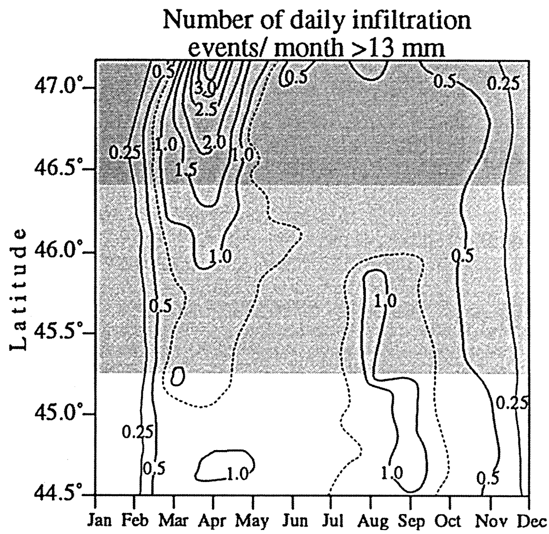

Fig. 7. Number of days per month with infiltration events larger than 13 mm (modelled), in study area averaged over the 1951 to 1992 period of record.

our findings here and in southern Michigan suggest that the thinner snowpacks that accumulate later in the winter, leading to lower amounts of snowmelt infiltration, may also be responsible for the weakly-developed character of soils in northern Minnesota. Similarly, in Finland, Jauhiainen (1973) found that soils at an interior location were

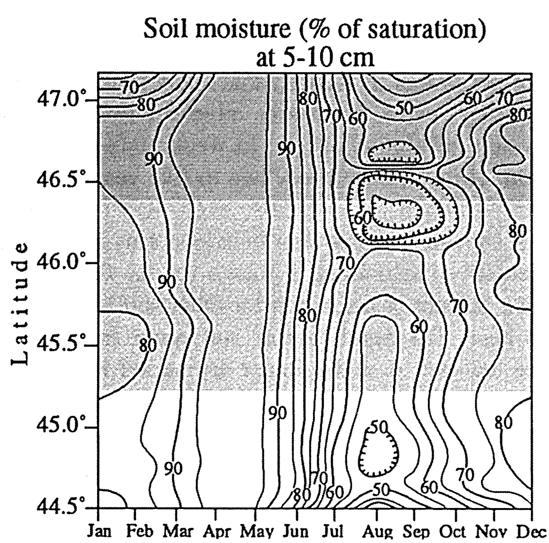

Fig. 8. Mean monthly soil moisture at 0.05–0.1 m (modelled) given as percentage of saturation in study area averaged for the 1951 to 1992 period of record.

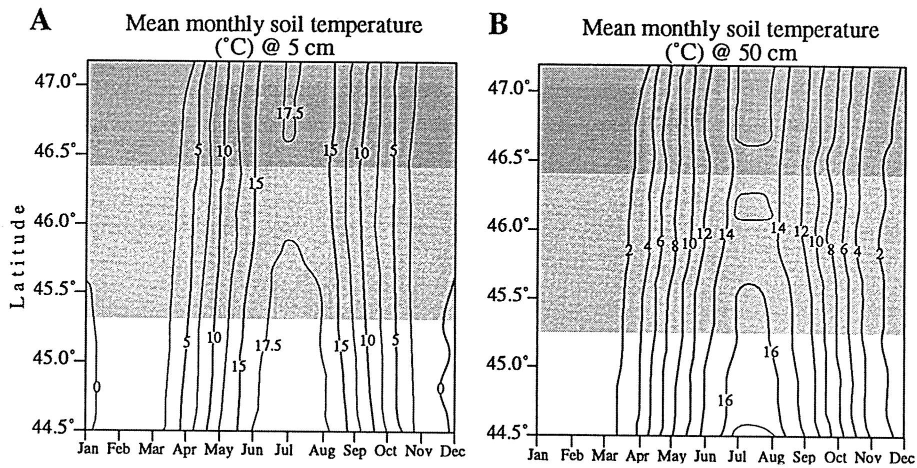

Fig. 9. Mean monthly soil temperature (modelled) for (A) 0.05 m depth and (B) 0.5 m depth in study area averaged for the 1951 to 1992 period of record.

more strongly podzolized than those near the coast. The interior location was snowier, having both deeper snowpacks and a greater percentage of its total precipitation falling as snow. Therefore, this study lends credence to, and provides an explanation for, a relationship that is commonly known by soil mappers in the upper midwest: spodic development increases in snowy areas.

4.3. *Climatic factors involved in podzolization: heat*

Because water flow through soil can depend on soil temperature, we created and examined maps of: (1) soil temperature, (2) freeze–thaw cycles, (3) length of freeze, and (4) runoff from frozen soil, and examined their spatial congruence with the regional-scale maps of podzolization.

Mean monthly soil temperatures for 0.05 and 0.5 m depths, as calculated by the hydrologic model, are presented in Fig. 9. Soil temperatures in the northern part of the study area are 1–2° colder than in the south, but only in the south are mean monthly temperatures $< 0°$, when winter snowpacks are thin or absent. Cooler overall soil temperatures in the north may favor podzolization by allowing for slower decomposition of organic detritus and forest litter. Indeed, the slowed (but not negligible) decomposition of the forest litter beneath deep snowpacks, coupled with long, continuous fluxes of meltwater, may be an important component of the podzolization process. Slow but continuous additions of organic chelates to the mineral soil may be facilitated by snowmelt-type infiltration, while the cooler soil temperatures, lasting well into early summer, may slow or negate microbial decomposition of the chelate complexes, allowing them to infiltrate deeply.

Soil freezing has been shown to be an important pedogenic process. It functions in the formation of soil structure and has an impact on soil nitrogen reactions, CO_2

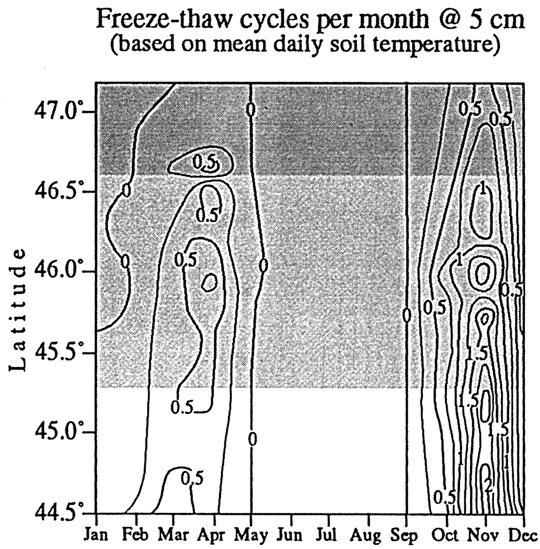

Fig. 10. Number of freeze–thaw cycles per month (modelled) defined by -1°C (lower) and 0°C (upper) soi temperature thresholds for 0.05 m soil depth. Calculated using average daily soil temperature.

evolution, and microbial populations (Wang and Bettany, 1993, 1994). Cates anc Keeney (1987) found large increases in N_2O in soil air immediately after a thaw, anc abundant evidence exists for a large pulses of N_2O from soils during the spring thav period. Freeze–thaw cycles, defined by -1° (lower) and 0°C (upper) temperature thresholds were calculated using the average daily soil temperature (Fig. 10). Modellec soil temperature data, calculated for each 20 minute time step, produced a similar patter to that shown in Fig. 10, suggesting that freeze–thaw activity is minimal in the stron podzolization zone and noticeably more common in the south, especially in November Soils in the strong podzolization also remain frozen for shorter periods of time than dc soils farther south (Fig. 9A, Fig. 11). This pattern is due to the higher incidence of early deep snowpacks in the north, which effectively insulate the soil from the cold air above (Wisconsin Statistical Reporting Service, 1970; Isard and Schaetzl, 1993). Soil fros diminishes in likelihood with depth (Fig. 11), and most often occurs in late autumn anc in spring, immediately following spring snowmelt. At these times, cold air combinec with thin or nonexistent snowpacks leads to ample heat loss from the upper layers of the soil. In mid-winter, where the snowpack is deep, the upward flux of ground heat fron deep soil layers melts any frost that may have formed in near-surface horizons, while the snowpack effectively insulates the soil from the cold air above.

Frozen soil is often impermeable to infiltration (Trimble et al., 1958; Burt anc Williams, 1976). For this reason, runoff (Fig. 12) is calculated by the model when wate inputs encounter frozen soil. (Because precipitation intensity data are not available runoff from summer thunderstorms cannot be computed by the model.) Runoff is temporally concentrated in March, April, November, and December, the months wher

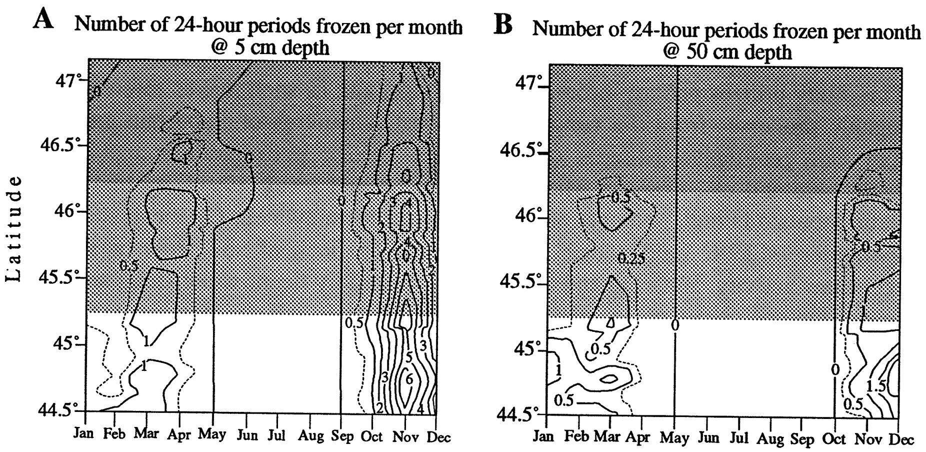

Fig. 11. Time per month that the soil is frozen in study area (modelled) averaged for 1951 to 1992 period of record: (A) Number of days (complete 24-hour periods) frozen at 0.05 m soil depth and, (B) Number of days frozen at 0.5 m soil depth.

soils in the study area are most likely to be both frozen (Fig. 11) and wet (Fig. 8). Precipitation falling (or snow melting) during these periods may not contribute to pedogenesis. Runoff from frozen soils is, therefore, most common in the central and southern part of the study area.

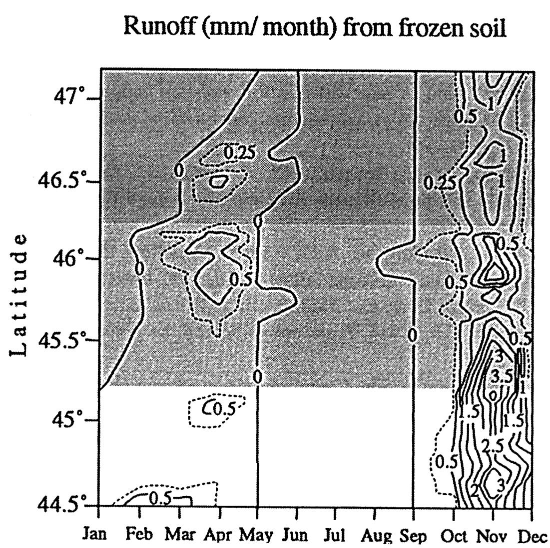

Fig. 12. Total monthly runoff (mm) per month from frozen soil (modelled) in study area averaged for the 1951 to 1992 period of record.

5. Conclusions

This study utilizes pattern–process relationships, and as such makes inferences about pedogenic processes from the congruence of spatial patterns of important soil climate factors and geographic patterns of Spodosols. Strong podzolization appears to be favored, here and elsewhere in Michigan (Schaetzl and Isard, 1991), by the following climatic inputs: (1) high amounts of infiltration during snowmelt, (2) mean summer soil temperatures $< 17°C$ (and cooler at depth), and (3) lack of soil frost or significant freeze–thaw activity. Both (1) and (3) are accentuated by deep, winter-long snowpacks. Climatic factors common to strong podzolization but which may or may not be important to the process include: (1) low soil water content during summer and (2) little, or only shallow, infiltration inputs during autumn.

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