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Pedology

Pulses of Podzolization: The Relative Importance of Spring Snowmelt, Summer Storms, and Fall Rains on Spodosol Development

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This study was performed in the eastern Upper Peninsula of Michigan, where Spodosols are extremely well developed. We instrumented a Typic Durorthod with zero-tension lysimeters to capture water leaving the O, E, and B horizons and with sensors to determine volumetric water contents with depth. We also occasionally measured snowpack depths and determined snow water equivalents. These field data were used to validate a hydrologic model that was run for the site using nearby National Weather Service (NWS) data. Good agreement between the modeled output and field data from the site enabled us to apply 1961 to 2013 NWS data from three additional stations along a transect that spans the range of podzolization strength in Michigan as inputs to the model. Soils remain dry throughout the summer and slowly wet up in fall. The more strongly developed soils in the north are slightly wetter in fall, facilitating breakdown of fresh litter and enhancing production of soluble organic materials. Their translocation into the mineral soil is presumably deepest and most pronounced during snowmelt, facilitated by a strong "pulse" of cold snowmelt water. This pulse comprises well over half of the annual flux of water at 100-cm depth, even though its timespan is short. Snowmelt fluxes are larger and of shorter duration in the north, where podzolization is strongest. By storing precipitation in a thick snowpack, the pedogenic system compresses inputs of water, creating deeper, more concentrated pulses of percolation when soluble organic materials are readily available; this is the essence of podzolization in this region.

Abbreviations: NSE, Nash–Sutcliffe efficiency statistic; NWS, National Weather Service; ODOE, optical density of the oxalate extract; SWD, snow water density; SWE, snow water equivalent; VWC, volumetric water content; ZTLs, zero-tension lysimeters.

One of the best ways to determine the most important drivers of a pedogenic process is to examine the state factors that come together in areas where that process is especially strong, or where the outcomes of that process, that is, soil morphologies, are particularly well expressed. Perhaps nowhere in the United States are spodic morphologies as strongly developed as in parts of Michigan's Upper Peninsula (Rourke et al., 1988; Witty, 1990; Schaetzl and Isard, 1996; Schaetzl, 2009) (Fig. 1). On uplands in this area, mixed forests on sandy parent materials, coupled with a cool, humid climate with heavy snows, work in concert to promote the deep leaching and acidic litter formation that foster intense podzolization (DeConinck, 1980; Lundström et al., 2000; Schaetzl and Harris, 2011). Here, Typic Haplorthods and Durorthods have exceptionally thick and white E horizons, often with deep tongues into the spodic (Bhs and Bs) hori-

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zons below (Fig. 2). These morphologies are indicative of strong podzolization. Particularly noteworthy is the fact that these soils have formed in only approximately 11,500 yr, since the end of the last glaciation (Johnson et al., 1997; Derouin et al., 2007; Blewett et al., 2009).

Previous work has shown that the intensity of podzolization across the Great Lakes region varies mainly as a function of climatic factors (Schaetzl and Isard, 1991, 1996; Mokma, 1991), or of climatic factors as they have (directly or indirectly) influenced vegetative cover and forest type (Mokma and Vance, 1989; Schaetzl, 2002). The character of forest vegetation does change across the Great Lakes region on a macro scale, from coniferous and mixed coniferous-deciduous forests in the north to broadleaf deciduous forests in the south (Curtis, 1959; Schaetzl and Isard, 2002; Harman, 2009) (Fig. 1c). However, because coniferous and mixed coniferous-deciduous forest vegetation favor podzolization, Spodosols are found almost exclusively north of the vegetative tension zone that separates these two vegetation zones (Fig. 1a and 1c). Thus, studies that evaluated and explained podzolization strength and character across the northern Great Lakes region have not normally controlled for vegetation (Schaetzl, 2002; Schaetzl et al., 2006).

Within the mixed coniferous-deciduous forest region, several studies have pointed to the important effect that snow, snowmelt, and the deep percolation that ensues have on podzolization (Schaetzl and Isard, 1991, 1996; Schaetzl, 2002). The theory can be represented as follows. Snowbelt areas, with their reliable and thick snow cover, tend to occur in belts that are 20 to 70 km inland, west and/or south of the Great Lakes (Eichenlaub, 1970; Norton and Bolsenga, 1993; Burnett et al., 2003; Andresen and Winkler, 2009; Henne and Hu, 2010) (Fig. 1b). In these areas, thick snows accumulate early, inhibiting soil freezing and keeping the soil and litter layers relatively warmer than in areas further inland which receive less snow (Isard and Schaetzl, 1993, 1995, 1998; Isard et al., 2007). As a result, soils in snowbelt areas stay unfrozen and permeable throughout the

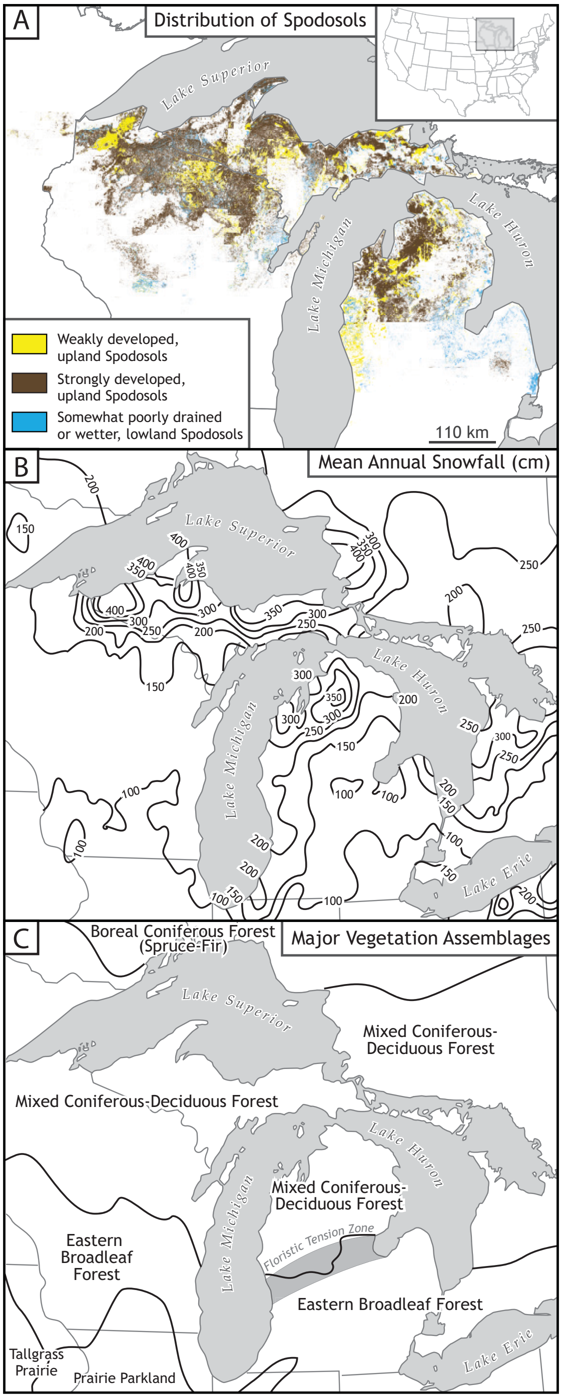

Fig. 1. Maps of environmental factors and soils (Spodosols) in the northern Great Lakes region. (a) Spodosols, as grouped into three general categories based on taxonomic subgroup (see below). Data source: NRCS SSURGO. (b) Mean annual snowfall 1951–1980. After Schaetzl and Isard (2002). (c) Major vegetation assemblages. After Schaetzl and Isard (2002). Further detail for Fig. 1a: Yellow (weakly developed, upland Spodosols): Entic Haplorthods and Aquentic Haplorthods. Brown (strongly developed, upland Spodosols): Alfic Fragiorthods, Alfic Haplorthods, Alfic Oxyaquic Fragiorthods, Alfic Oxyaquic Haplorthods, Aqualfic Haplorthods, Fragic Haplorthods, Lamellic Haplorthods, Lithic Haplorthods, Oxyaquic Fragiorthods, Oxyaquic Haplorthods, Oxyaquic Ultic Haplorthods, Typic Durorthods, Typic Fragiorthods, and Typic Haplorthods. Blue (somewhat poorly drained or wetter, lowland Spodosols): Aeric Haplaquods, Alfic Epiaquods, Alfic Haplaquods, Argic Endoaquods, Argic Fragiaquods, Entic Haplaquods, Histic Epiaquods, Typic Duraquods, Typic Endoaquods, Typic Epiaquods, Typic Fragiaquods, Typic Haplaquods, Typic Haplorthods, and Ultic Fragiorthods.

Fig. 2. Soils (Typic Haplorthods and Durorthods) exposed in a trench at the Newberry study site before sampling and installation of the field equipment. Note the prominent tongueing that is typical of these soils. Tape scale in centimeters.

winter. In spring, large pulses of snowmelt water can freely infiltrate into and percolate through these soils, often as nearly continuous, several daylong wetting events that penetrate the solum (Schaetzl and Isard, 1991, 1996). The deep, reliable, and continuous snowpacks in snowbelt areas also insulate the fresh litter in the O horizon, facilitating its steady breakdown throughout the winter, promoting the production of soluble organic acids (Taylor and Jones, 1990; Brooks et al., 1996). The wetting events that occur during snowmelt can readily translocate these soluble organic acids from the O horizon, deep into the mineral soil below. Litter in snowbelt areas also stays wetter longer into spring, which is the main fire season in the Great Lakes area. Thus, the litter layer is destroyed less often by fire in snowbelt areas, maintaining the source of organic acids and, hence, further facilitating podzolization. Areas outside of snowbelt areas, where fire is more common, have thinner O horizons and less developed soils (Mokma and Vance, 1989; Schaetzl, 2002). In summary, in areas of weaker podzolization, fire is more prevalent, and hence litter is thinner and soil frost is more likely. As a result, less water is available for springtime percolation because of thinner snowpacks and more frequently frozen soil. Much of the little snowmelt that exists runs off and does not participate in pedogenesis.

The theory described above was first hypothesized from observations of spatial co-association, that is, spodic development is stronger in snowbelt areas (Jauhiainen, 1973), and later validated by the soil hydrology and temperature model of Schaetzl and Isard (1991, 1996). This model used as inputs NWS data from several stations in the northern Great Lakes region, producing output on soil temperatures, snowpack thicknesses, and percolation for sites within and outside the snowbelt. Percolation was modeled for various depths, but was not validated with field data. Their internally developed model showed that Spodosols are dominant and more strongly developed in areas where soil freezing is minimal, where snowpacks are thickest, and where snowmelt percolation is greatest. They argued that it is not only the amount of percolation that matters, but that the slow, steady flux of cold water during snowmelt is particularly effective at translocation of organometallic complexes. Their modeled data showed that snowmelt fluxes of water into the mineral soil are only one third as large in interior, nonsnowbelt areas. Interestingly, their modeled data also showed that a small, secondary pulse of percolation, which occurs in autumn, is found in the nonsnowbelt areas, but is commonly absent or much smaller in snowbelts.

Although the models results provided by Schaetzl and Isard (1991, 1996) validate the "snowbelt-podzolization theory" for the Great Lakes region, the aim of this paper is to add to the science, report new data, and build on this theory in two ways:

1. By using field data from our research site as a means of model validation, we subsequently employ a widely used hydrologic model to generate data on soil water flux, as well as soil and snowpack water contents.

2. We evaluate not only the effects of spring snowmelt vs. fall rains on percolation and pedogenesis, as was done previously, but also examine and compare these data with data on percolation as driven by summer rain events.

The overall purpose of our research is to test and evaluate the effect of thick snow and snowbelt conditions on podzolization along a transect of four sites in Michigan from a snowbelt core to an area at the far southern margin of the podzolization regime. All sites are within the mixed coniferous-deciduous forest region, that is, the influence of vegetation is generally controlled for. Our findings will add considerably to the discourse on the drivers of podzolization, as well as inform the hydrologic and forestry communities about the effects of snowmelt, and fall and summer rains, on deep soil wetting and groundwater recharge.

MATERIALS AND METHODS

Study Area and Research Sites

We report on modeled, hydrologic data for four NWS sites (Newberry, Gaylord, Houghton Lake, and Mt. Pleasant) that form a north-south transect across the podzolization region of northern Michigan (Fig. 3). Along the transect, snowfall totals and podzolization intensity generally decrease, and the climate warms. The three northernmost sites are in the mixed coniferous-deciduous forest zone, whereas Mt. Pleasant is located within the floristic tension zone (Fig. 1c). The site near Newberry, in Michigan's Upper Peninsula, represents a Lake Superior snowbelt location that has, on uplands, strongly developed Haplorthods with thick, bright E horizons and ortstein-rich spodic horizons (Fig. 2). Gaylord is also in a snowbelt, but one from Lake Michigan. Even though Gaylord receives more snow than Newberry (Fig. 3), we chose it because it often has more wintertime melt events and, as a result, usually enters the snowmelt season with less snow on the ground. Upland soils on sandy sites near Gaylord are Typic and Entic Haplorthods, and ortstein is uncommon (Mokma and Vance, 1989; Schaetzl, 2002; Mikesell et al., 2004; Henne et al., 2007). Some of the upland, sandy soils near Houghton Lake are Entic Haplorthods, although most are Lamellic Udipsamments, suggestive of increasingly weakening podzolization. Spodosols are not present on uplands near Mt. Pleasant; sandy uplands are mapped as Lamellic Udipsamments or Arenic Hapludalfs. Some of these soils have reddened B horizons, indicative of in situ weathering, akin to the traditional concept of braunification (Cointepas, 1967; van Breeman and Buurman, 1998; Chen et al., 2001), that is, very weak podzolization. E horizons in sandy soils at this latitude in southern Michigan are thin, at best, and often absent (Arbogast et al., 1997).

Soil sampling and installation of field equipment were performed in May 2012 at a wooded site 36 km east of the Newberry NWS station and 2.9 km SSW of the intersection of Highways 28 and 123 (Fig. 3). It is on a sandy upland known locally as Hulbert Island (Schaetzl et al., 2013), for its location within Glacial Lake Algonquin (Karrow, 2004; Drzyzga et al., 2012). Soil parent materials are deep, well-sorted outwash sands, probably associated with a former retreating ice margin. The site is mapped as Kalksaka sand (Sandy, isotic, frigid Typic Haplorthods) by the

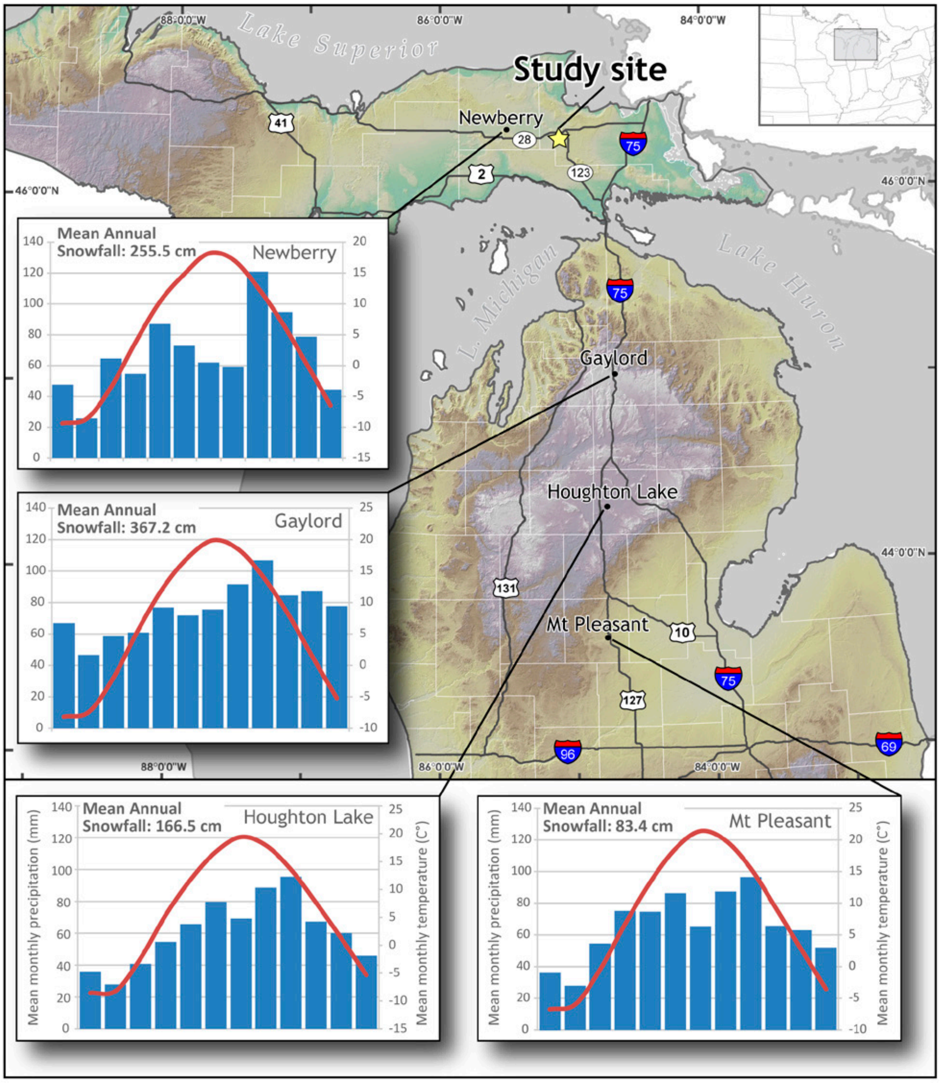

Fig. 3. Locations of the four National Weather Service (NWS) stations examined in this study along with their climographs. The study site, where the Wallace soil was sampled and where the field instrumentation is installed, is east of Newberry.

NRCS (Whitney, 1992). Vegetation at the site is dominated by sugar maple (Acer saccharum), with minor components of red maple (Acer rubrum), black cherry (Prunus serotina), and American beech (Fagus grandifolia). Presettlement vegetation in the region was mapped by the General Land Office survey as beech– sugar maple–hemlock (Tsuga canadensis) forest (Comer et al., 1995).

Field Methods

At the Newberry site, we excavated a 1.8-m-deep, 6-m-long trench by backhoe (Fig. 2) on a level area (2% slope) that lacked evidence of prior disturbance by tree uprooting (Šamonil et al., 2013). To avoid disturbing or compacting the soil on the uphill side of the trench, we covered it with 3/4-inch plywood sheeting during excavation and sampling. The undisturbed face was cleaned, and a pedon described and sampled, using standard methods (Schoeneberger et al., 2012).

Zero-tension lysimeters (ZTLs) were installed at the base of the O, E, and Bs horizons at depths of 9, 35, and 59 cm, respectively (Fig. 4). This type of lysimeter is particularly effective at capturing water percolating rapidly as saturated flow (Toosi et al., 2014). The lysimeters were 16-cm-diameter high-density polyethylene funnels filled with combusted (550°C), acid-washed (10% HCl) quartz sand, as described by MacDonald et al. (1993) and Wilson et al. (1994). They were installed approximately 25 cm into the face of the trench at staggered locations (Fig. 4c). Funnels were supported with custom-made ABS plastic holders and outfitted with four turnbuckles that were used to raise the lysimeters into tight contact with undisturbed soil above. Drainage water was carried by PTFE tubing, by gravity, from the ZTLs to a 4-L glass collection bottle at the bottom of the pit, with a separate PTFE tube running from the collection bottle to the soil surface (Fig. 4b).

Volumetric water sensors (5TM, Decagon Devices, Pullman, WA) were installed into the pit face at predetermined depths of 5, 20, and 45 cm below the mineral soil surface and connected to a Decagon EM50 datalogger which recorded soil volumetric water content (VWC), which we estimated by using factory default calibration at hourly intervals. We used factory default settings for our sensors because under soil conditions of our site, soil-specific calibration vs. the factory default calibration should only change VWC output by 1% or less (Technical Support, Decagon Devices, personal communication, 2014).

Before backfilling the pit, we installed a sheet of 0.15-mm-thick polyethylene film against the exposed pit face to isolate it from the fill materials in the pit. Because the VWC-sensing equipment is part of a larger project in which several additional pedons are being monitored, the sensors were placed at standard depths, which in most cases do not coincide exactly with the horizon-based placement of the ZTLs.

Lysimeter collection bottles were initially cleared of any accumulated liquid in late June 2012; from that point forward, leachate was collected a few days after any rain or snowmelt event of >2-cm water equivalency. Event-based sampling decisions were based on precipitation models from the NWS Advance Hydrologic Prediction Service (http://water.weather.gov/precip/), and snowmelt models from the NWS National Operational Hydrologic Remote Sensing Center (http://www.

Fig. 4. Photos showing the characteristics and field installation of the lysimeters at the Newberry site. (a) Close-up of a zero-tension lysimeter, installed (in this case) to capture percolation from the O horizon. Note the tube that carries the leachate to a lower-lying glass bottle. (b) Installed lysimeters, draining to a glass bottle array, partially buried in sand in a plastic box at the base of the soil pit. The bottles are used to capture and store the leachate. The gray case holds the hose and electrical leads from the subsurface, facilitating the sampling of soil solutions and the downloading of data after the pit has been backfilled. (c) A completed array of zero-tension lysimeters, ready for burial.

nohrsc.noaa.gov/nsa/). Each of the soil water collection bottles was evacuated with a handheld pump, and the volume of leachate collected was recorded. We were careful to never walk on or disturb the soil surface that spanned the area from the pit face backward for 4 m, that is, the area formerly covered by the plywood.

We monitored the development, water content, and melting of the snowpack at the study site approximately every 2 to 3 wk during the winters of 2012–2013 and 2013–2014, and more often during the snowmelt period. At each visit we measured snow depth at three locations within the 6- by 4-m plot above the lysimeters and VWC sensors. We also collected snowtube samples from two undisturbed locations within 25 m of the site, which were used to determine snow water density (SWD) in cm water cm^{-1} snow depth. Average SWD at each sampling date was multiplied by snow depths to calculate snow water equivalent (SWE) (cm).

Laboratory Methods

All soil samples were air dried and lightly ground to pass a 2-mm sieve. For particle size analysis of A and Bhs samples, organic matter was removed using dilute (15%) H_2O_2. Two gram subsamples were dispersed in a water-based solution of $(NaPO_3)_{13}\times Na_2O$, and then, after shaking for 2 h, run on a Malvern Mastersizer 2000E laser particle size analyzer (Miller and Schaetzl, 2012). Soil pH was analyzed using a glass electrode in a 2:1 water/soil mixture. Samples from each horizon were pulverized in a ball mill and analyzed for total C by dry combustion gas chromatography on a Costech ECS 4010 (Costech Analytical Technologies, Valencia, CA).

Iron (Fe), aluminum (Al), and silicon (Si) extractions were conducted on all mineral soil samples by using traditional methods. A sodium citrate-dithionite solution was used to extract amorphous and crystalline (commonly referred to as "free") forms of Fe, Al, and Si (Mehra and Jackson, 1960). Organically bound forms of Fe, Al, and Si, which provide a good indicator of the abundance of organometallic complexes, were extracted with Na-pyrophosphate (McKeague, 1967). A third extractant, acid ammonium oxalate, was used to obtain the amounts of Al and Fe from organic complexes, as well as from poorly crystalline minerals such as ferrihydrite, allophane, and imogolite (Daly, 1982). Finally, the optical density of the oxalate extract (ODOE) was determined on a UV spectrophotometer at 430 nm. Values for ODOE are important for classification purposes (Soil Survey Staff, 2010).

Modeling Methods

We used the BROOK90 hydrologic model (Federer, 2012) to model snowpack dynamics and soil water fluxes for soils near the four NWS station discussed above. BROOK90 has been applied to temperate forest ecosystems around the world (Holst et al., 2010; Bencokova et al., 2011). It is particularly suited to our research because it was developed and initially calibrated for the Hubbard Brook watershed in New Hampshire, which has climate and forest cover similar to northern Michigan (Lawrence et al., 1995; Fisk et al., 2002; Jost et al., 2005; Beckers et al., 2009). Another key benefit of BROOK90 is that it allows for the modeling of water flow within different soil layers, rather than treating soil as a single volume; this characteristic is essential to our hypotheses about percolation depths. The model uses as inputs daily temperature (high and low) and precipitation (water equivalent) data, which we obtained for NWS stations at Newberry, Gaylord, Houghton Lake, and Mt. Pleasant. Missing data, for example, daily precipitation totals or temperature readings, in these files were cut and pasted from the nearest available NWS station or stations.

Before extrapolating the model to historical NWS data, we used our field data on snowpack dynamics, soil VWC, and hydrologic fluxes to assess the accuracy of the model. We used the default "Temperate Deciduous Forest" canopy input file and "Top Down" flow input file. We then modified the default "Location" and "Soil" input files to incorporate the soil and site attributes measured in the field and laboratory, that is, slope gradient and aspect, and soil horizon thickness and texture. We input daily precipitation and temperature data to the model from the NWS weather station at Newberry. We began our model calibration runs on 1 Apr. 2012 to allow for a 3-mo "spin-up" period before our first field lysimeter collection began on 28 June 2012 and extended through two full winters, ending on 12 May 2014.

We began by comparing (i) modeled daily SWE against that measured in the field, (ii) modeled soil VWC at depths of 5, 20, and 45 cm against daily average VWC from the soil moisture sensors at equivalent depths, and (iii) modeled vertical matrix flow at 0, 26, and 50 cm against water collected at the same depths by the ZTLs. By using default parameters, the BROOK90 model provided an excellent fit to field-collected data on soil VWC and water fluxes during the snow-free season, but underestimated the amount water contained in the snowpack when compared with field data. We were confident that this discrepancy was not due to errors in field measurement of SWE, because our field data aligned well with the NWS's modeled data for SWE (http://www.nohrsc.noaa.gov/nsa/). Thus, we iteratively adjusted the following model parameters related to snowpack dynamics: RSTEMP (base temperature for separation of precipitation into rain and snow), KSNVP (correction factor for snow evaporation), GRDMLT (constant rate of groundmelt at the base of the snowpack), and FSINTS (fraction of snowfall intercepted per projected stem area). We made the largest changes to RSTEMP because inspection of internal flows within the model showed that the model was assigning too great a proportion of wintertime precipitation as rain. We also set GRDMLT to zero, based on the fact that our uppermost zero-tension lysimeters remained dry for the entire winter, except for during discrete warming events. After adjusting the model to more accurately capture snowpack dynamics we reran the model for the June 2012 to May 2014 period and compared model output against field-collected VWC and lysimeter water flux to assess accuracy of the final model configuration.

After the calibration-validation phase, we ran the model by using data for the four NWS stations for the years 1961 to 2013, but discarding data for 1961 because we considered it a spin-up year. This setup left us with 52 yr of viable data for each of the four sites, for which we determined daily values of water flux at four depths (9, 35, 59, and 100 cm), and VWC at two depths (10 and 100 cm), representing shallow (10 cm) and deep (100 cm) percolation events. These modeled data were summarized and examined with respect to the snowmelt podzolization hypothesis. BROOK90 does not output data on soil temperature.

MAPPING METHODS

To better evaluate podzolization strength, as it covaries with climatic state factors across the region, we first downloaded the SSURGO data for Minnesota, Wisconsin, and Michigan and combined and rasterized the files in a GIS. For the 232 Spodosol series that are mapped in the region, we then determined their drainage classes by examining official series descriptions on the NRCS website (https://soilseries.sc.egov.usda.gov/osdname.asp). For each series that was moderately well-drained or drier, we determined its POD Index, which is a field-based index of podzolization strength (Schaetzl and Mokma, 1988) that has been applied successfully in a variety of environments where only morphological data are available (Anderton and Loope, 1995; Goldin and Edmonds, 1996; Arbogast and Jameson, 1998; Wilson, 2001). We did not assess strength of development for the somewhat poorly drained Spodosols because podzolization in the presence of a high water table is typically a function of water table dynamics and chemistry, rather than regionally varying climatic and vegetative factors (Harris, 2000; Schaetzl and Harris, 2011). The POD Index uses as inputs the color hues and values of the E and B horizons, and assesses their relative differences. Larger numbers of B horizons, with increasing contrast in color hue and value relative to the E horizon, work in tandem to create larger POD Index values. POD Index data were joined to the attribute table of the soils in the GIS. We next (randomly) assigned 5000 data points across the three state region, but only in the area where Spodosols occur (Fig. 1). At the 1438 locations where the point intersected an upland Spodosol map unit, we determined the POD Index from the SSURGO database and kriged the point data in ArcGIS 10.1 to arrive at a smoothed isoline map of POD Index values across the study area (Fig. 5).

RESULTS AND DISCUSSION

Soil Characterization

The sampled pedon near the Newberry NWS station, mapped within the Kalkaska series (Sandy, isotic, frigid Typic Haplorthods), is an extremely well-developed Spodosol, with deep E and B horizon tongues and continuous ortstein (Fig. 2). Ortstein was less continuous, but common, in lower B horizons (Table 1). All horizons except the Bhs horizon (loamy sand) were sand textured, with medium and fine sands dominating (Table 1). Clay contents were less than 2.5% for all horizons, and no clay was present in the BC horizon, indicating the well-sorted nature of the outwash sand parent materials here. Coarse fragments were observed only below the solum.

Both morphological and chemical data confirm that the pedon sampled near Newberry illustrate the exceptional development of this pedon, as is typical for this area (Franzmeier and Whiteside, 1963; Messenger et al., 1972; Barrett and Schaetzl, 1992) (Fig. 5 and Table 2). The upper B horizon has very high amounts of extractable Fe, Al, and Si, both in "free" (dithionite-extractable) and "organically bound" (pyrophosphate-extractable) forms. Similarly, the E horizon has very low amounts of these same elements (Fig. 6a and 6b and Table 2). Collectively, these data are indicative of strong profile differentiation via podzolization. The pedon's POD Index of 31 is unusually high; most well-developed, upland Spodosols have POD Indices between 6 and 15 (Schaetzl and Mokma, 1988), and most Spodosols across the lake states region have POD Indices <6 (Fig. 5). Chemical data for this pedon are equally impressive. Data for ODOE, used to identify spodic materials, illustrate the strong development of this pedon. Values for ODOE for the B horizon must be at least 0.5 and twice as large as the E horizon to qualify as spodic materials (with some exceptions; see Soil Survey Staff, 2010). The ODOE for the Bhs horizon for this pedon is 1.16, while the E horizon has an ODOE of 0.01. This large difference illustrates the magnitude of podzolization in this soil. On the basis of ODOE and horizon color data, the B horizon easily meets the criteria for spodic materials (Soil Survey Staff, 2010). The presence of ortstein enables the soil to be classified as a Typic Durorthod, within the Wallace series (Sandy, mixed, frigid, shallow, ortstein Typic Durorthods). Wallace soils are some of the best developed Spodosols in Michigan.

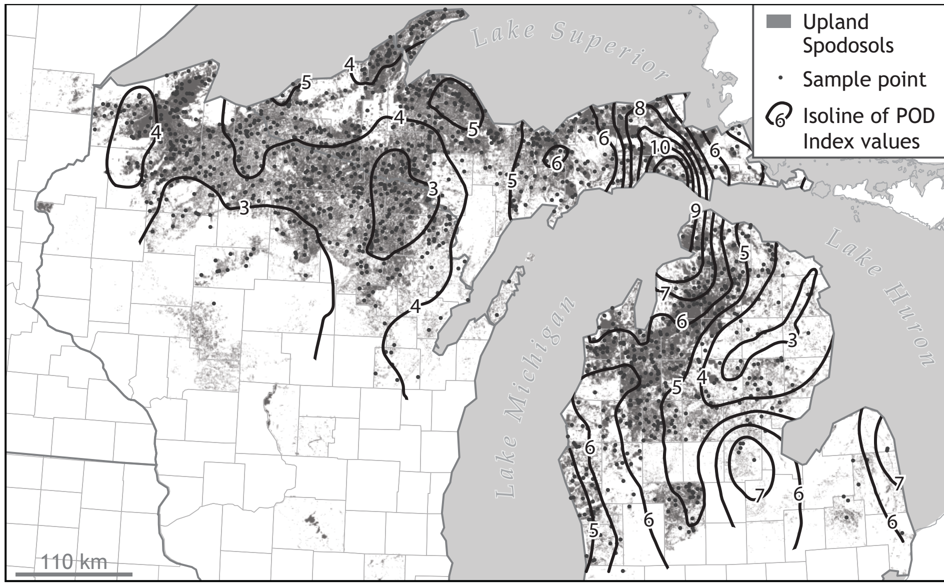

Fig. 5. Isoline map of interpolated POD Indices for upland Spodosols in the northern Great Lakes region. Higher values indicate stronger soil development. See text for details on the mapping procedures.

Soil extraction data (Table 2) suggest that podzolization is active and ongoing in the Newberry pedon. Podzolization, as driven by organic acid–mediated processes, for example, cheluviation or the fulvate-complex theory, is indicated by high levels of dithionite- (Fig. 6a) and pyrophosphate-extractable (Fig. 6b) Fe and Al, relative to those same extractions from the E horizon (Sauer et al., 2007). Also indicative of this process are the high ratios of organically bound Fe and Al to amorphous Fe and Al, that is, (Alp + Fep)/(Ald + Fed) in the upper profile (Fig. 6c). Peaks of all these kinds of data in the upper B horizon suggest that metal cations are being actively translocated to the B horizon as chelate complexes. Sauer et al. (2007) suggested that this form of podzolization is typical of soils formed on "poor sands" (cf. Wang et al., 1986). Evidence also exists for an inorganic podzolization pathway in this soil, as shown in Fig. 6c, which illustrates high levels of inorganically bound Fe, Al, and Si, mainly in the lower profile. These depth plots may imply that metal chelates that have been translocated to the B horizon become decomposed, facilitating the formation of inorganic metal sols, which are then free to translocate. Deeper translocation of Al than Fe, as is suggested here, is common in strongly developed Spodosols (Lundström, 1993). Regardless of the specific pedogenic mechanism(s) in-

Table 1. Physical and morphological characterization data for the soil at the Newberry site.

Horizon	Depth	Color Moist	Structure†	Texture Class	Bulk Density	Boundary (distinctness, topography)	Comments
	cm				g/cm^3		
Oe	0–7	n/a	n/a	n/a	not determined	clear, smooth	Some Oi and Oe material also present
E	7–23	7.5YR 6/3	w, f-m, sbk	sand	1.18	clear, irregular	
Bhs	23–34	2.5YR 2.5/2	w, f-m, sbk	loamy sand	1.08	clear, broken	»100% ortstein
Bsm	34–47	2.5YR 2.5/3	cemented	sand	1.36	gradual, broken	»20% ortstein
Bs1	47–62	7.5YR 4/6	w, m-c, sbk	sand	1.37	diffuse, broken	
Bs2	62–82	10YR 5/6	loose	sand	1.56	diffuse, irregular	
BC	82–130+	10YR 5/4	loose	sand	1.60		Few, thin (1 mm) clay lamellae; 2% coarse fragments

† w, weak; f-m, fine and medium; sbk, subangular blocky; m-c, medium and coarse.

Table 2. Chemical characterization data for the soil at the Newberry site.

Horizon	Organic C	pH	Ammonium Oxalate Extractions (Fe, Al, Si)	ODOE†
	%	2:1 water/soil	ug/g	
Oi	46.7	–	n/a	n/a
Oe	42.0	–	n/a	n/a
E	0.1	4.5	44, 55, 1	0.01
Bhs	1.2	4.4	7739, 3802, 309	1.16
Bsm	0.7	4.8	4551, 5988, 990	0.67
Bs1	0.3	5.0	1108, 2783, 704	0.14
Bs2	0.3	5.0	1043, 3076, 852	0.13
BC	0.0	5.2	272, 558, 146	0.04

† Optical Density of the Oxalate Extract. See also Fig. 6 for other chemical characterization data.

volved, these data illustrate the strong development of the soil at the Newberry site and point to the importance of deep translocation processes, driven by percolating water, in its genesis.

Hydrologic Model Performance and Types of Output

Our final model configuration, with the following new parameter settings, RSTEMP = 3.0, KSNVP = 2, GRDMLT = 0, and FSINTS = 0, produced an excellent match to field data for snowpack dynamics (Fig. 7). In particular, the calibrated model did an excellent job of predicting the peak storage of water in the snowpack, as well as the timing of the spring melt and the final loss of the snowpack. The model only slightly underestimates SWE at the research site, probably because the meteorological data are from the city of Newberry, which receives slightly less snow than the instrumented field site. Most importantly, the modeled SWE data align well by winter's end; the model melts the pack at the same time and rate as observed in the field (Fig. 7).

Figure 8 compares daily VWC from our final model configuration against raw VWC data recorded by soil moisture sensors in the field. Whereas the model accurately resolves the depth, timing, and relative magnitude of wetting events at the Newberry site, offsets in absolute values of VWC between sensors and the model occurred; these appear to be stable over the study period. The uppermost two sensors registered VWC values that were consistently higher than the BROOK 90 output, whereas the deepest sensor registered VWC data that were consistently lower. The cause of the consistent offset in absolute VWC values is unclear; they may be arising from variability in the local soil environment around the sensors in factors such as contact with particulate organic matter or roots, contact with the soil matrix, or proximity to preferential flow paths. Interestingly, the offset was greatest in the topmost layer where particulate organic matter and root density should be highest.

To address this discrepancy, we applied a constant correction factor to our sensor data, based on the average difference between sensor values and model output (−0.067, −0.020, and +0.019 m^3/m^3 for 10-, 25-, and 50-cm sensors, respectively) (Fig. 8d, 8e, and 8f). We then calculated the Nash–Sutcliffe efficiency statistic (NSE) to evaluate model performance, using both raw and offset-corrected data. Values of NSE can range from −¥ to 1, with values greater than 0 typically viewed as acceptable (Moriasi et al., 2007). Nash–Sutcliffe efficiency statistics for raw data (Fig. 8a, 8b, and 8c) were all below 0 (−4 to −0.06), indicating unac-

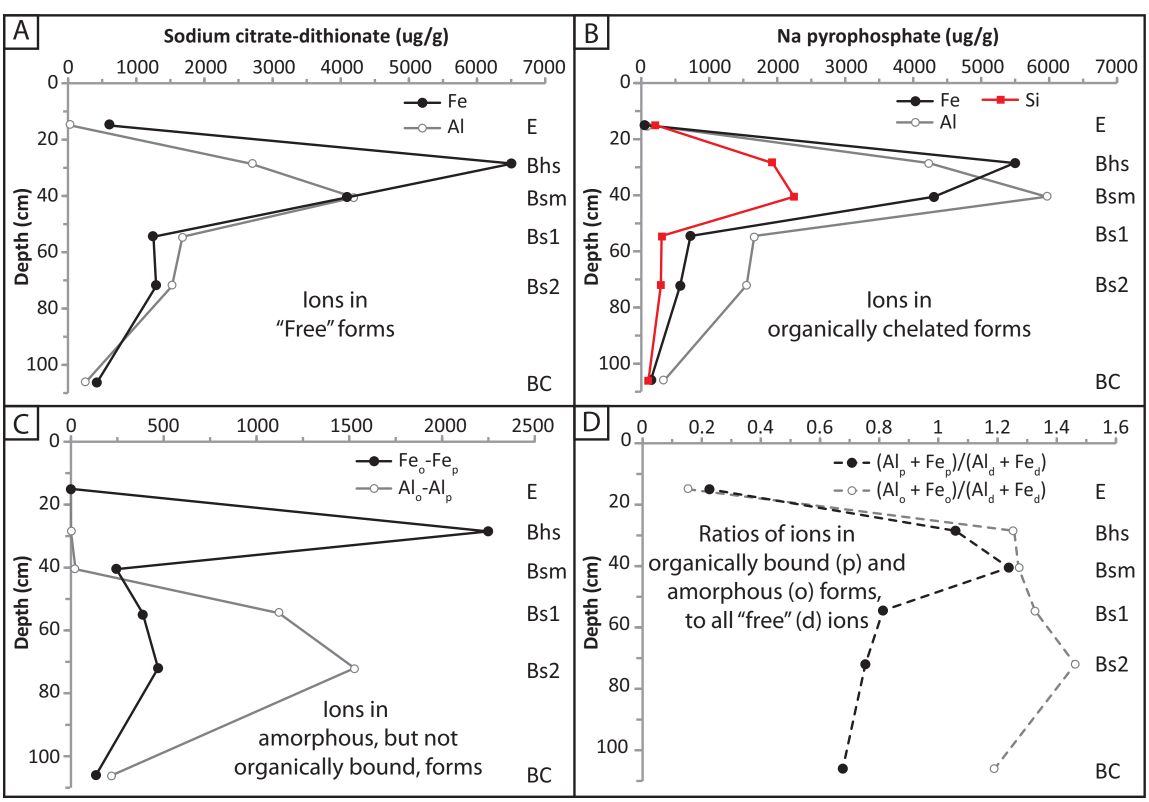

Fig. 6. Depth plots for various chemical data for the Newberry pedon. See also Table 2 for additional data.

ceptable model performance. whereas NSEs for offset-corrected data ranged from 0.64 to 0.93, clearly demonstrating that BROOK90 is able to accurately predict the timing and depth of soil wetting and drying at our sites.

We argue that offsets in the absolute value of soil moisture do not necessarily point to poor performance of BROOK90 at simulating water fluxes in our study site, in part because we found good agreement between measured vs. modeled flux of water at the different soil depths (Fig. 9). In this case, NSEs for absolute values of measured water flux vs. modeled water flux were all >0, indicating satisfactory model performance (NSE = 0.37, 0.37, and 0.59 for the 9-, 35-, and 59-cm depths, respectively).

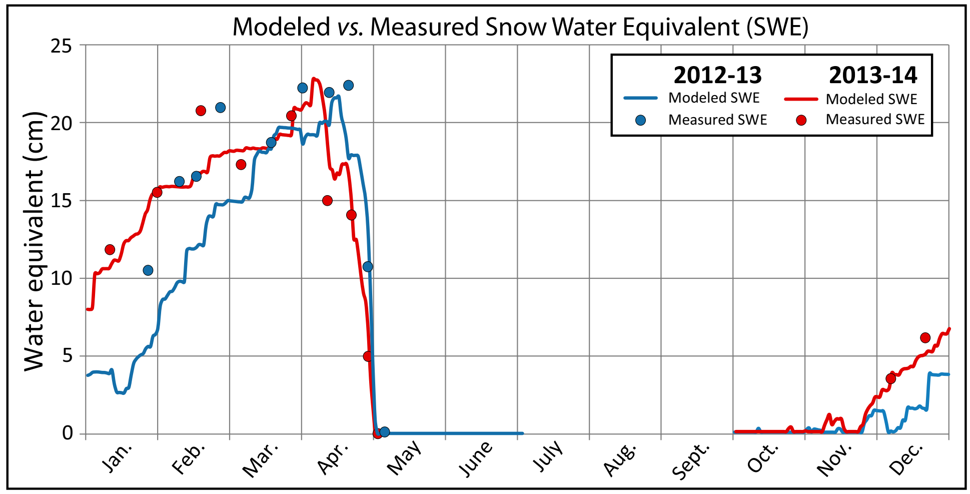

Fig. 7. Modeled vs. measured snow water equivalent (SWE) (snowpack) data for the Newberry study site for the 2012–2013 and 2013–2014 winters.

Note that during periods of high flux, water flux captured by the lysimeters was usually higher than that estimated by the model. In fact, at many of these times we captured more water in our lysimeters than actually entered the system via precipitation (data not shown). We attribute this discrepancy to the fact that the support frames surrounding the lysimeters (Fig. 4a) created an impervious surface, possibly channeling additional water into the lysimeter during periods of saturated flow. Because we are comparing relative amounts of flux and VWC among the four stations, the absolute differences shown in Fig. 8 and 9 are not critical to this analysis.

Snow and Soil Hydrology: Intersite Comparisons

Both modeled and measured SWE data for the Newberry site show the steady buildup of the snowpack during the winters of 2012–13 and 2013–14, followed by rapid loss during a brief and often intense snowmelt period (Fig. 7). Because of the agreement between the modeled and field-generated SWE data, we applied the model to the 1961–2013 data from the four NWS stations to examine interannual snowpack dynamics across the climatic transect (Table 3). To better visualize the intersite differences in SWE throughout the year, we plotted only the 52-yr means (Fig. 10).

All four sites exhibit steady snowpack accumulations throughout the winter, followed by a fairly rapid melt period (Fig. 10). Snowmelt tended to start slowly and then become more rapid later in spring. The mean snowmelt period (from peak snowpack to SWE = 0) ranges from 43 (Mt. Pleasant) to 58 d (Gaylord) in length. The two northernmost sites (Newberry and Gaylord) normally have much deeper snowpacks and greater SWEs than do Houghton Lake and Mt. Pleasant (Table 3). However, the northern sites also have higher rates of loss of water from the snowpack, suggesting that the flux of snowmelt water

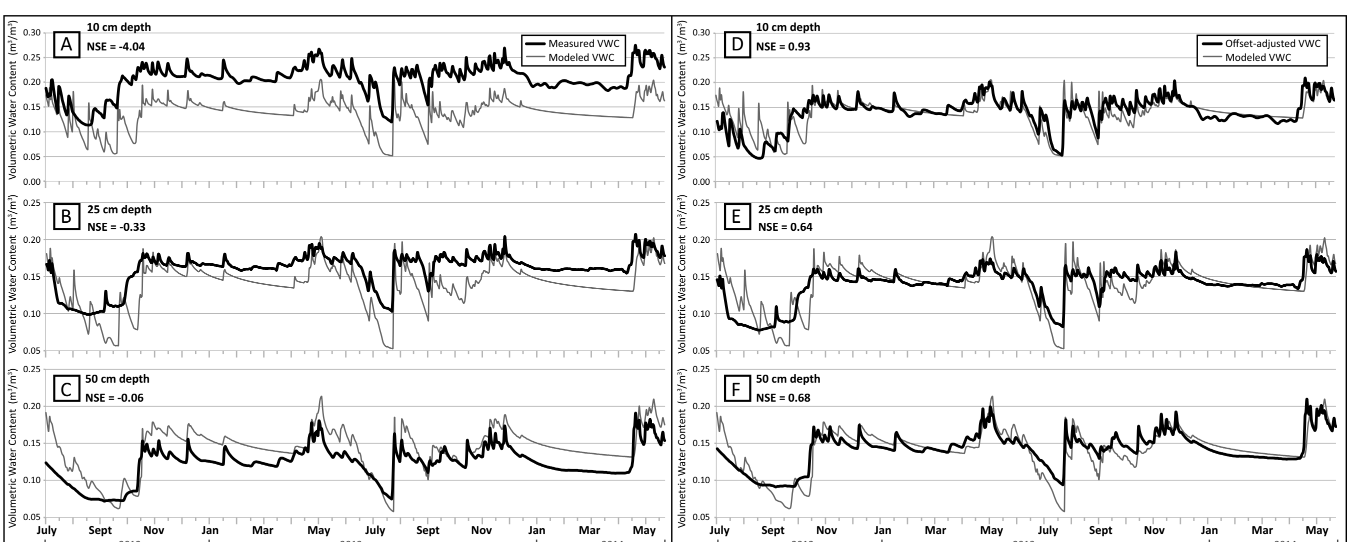

Fig. 8. Modeled vs. measured volumetric water contents (VWC), for three different depths, for the Newberry study site from 1 July 2012 through 12 May 2014.

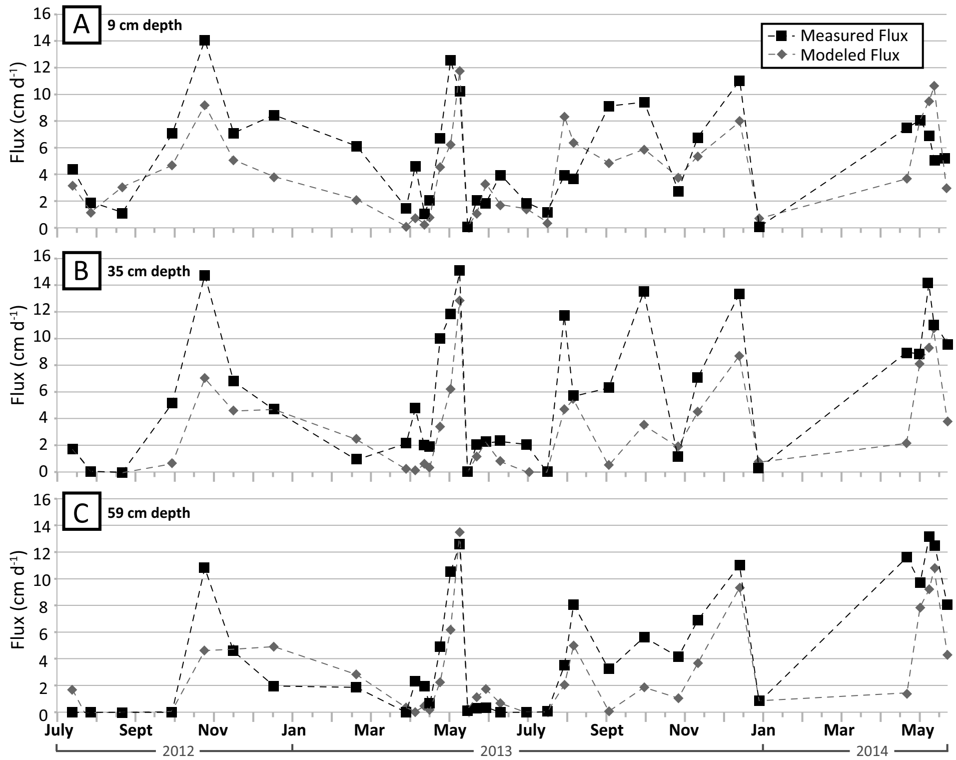

Fig. 9. Modeled vs. measured water flux at the three depths of measurement for the Newberry study site from 1 July 2012 through 12 May 2014.

into the soil is faster and larger here in this area of strong podzolization. In the north, thicker snowpacks, which are also more consistently present throughout the winter (Bolsenga, 1967) (Table 3), also inhibit deep soil freezing, allowing soils here to remain permeable throughout the snowmelt period (Schaetzl and Isard, 1991; Isard and Schaetzl, 1995, 1998). Soil temperature data for the Newberry pedon (see supplementary data) confirm that the upper profile of the mineral soil cools down to 1 to 3°C in winter, beneath thick snowpacks, but does not freeze. Farther south, thinner snowpacks can lead to frozen soil, which promotes runoff of snowmelt water (Schaetzl and Tomczak, 2002), doing little to promote podzolization. In summary, at sites father north, where snowpacks are thicker and more reliable, a larger and more rapid pulse of snowmelt water occurs.

Soil Water Contents and Fluxes: Intersite Comparisons

Directly and indirectly, water drives podzolization (DeConinck, 1980; Buurman and van Reeuwijk, 1984; Van Breeman and Buurman, 1998; Lundström et al., 2000). A udic soil moisture regime helps sustain the forest vegetation that produces the acidic litter necessary for the process, while also facilitating weathering of minerals and decay of O horizon materials. Abundant (and deeply) percolating water is the driving force behind translocation of soluble materials, leading to the strongly differentiated profile that typifies Spodosols (Fig. 2). In this section, we focus both on soil wetness and water content, and end with a discussion of how soil water fluxes vary with depth and over time.

The four transect sites exhibit considerable seasonal variation in soil water contents, as would be expected for the area: (i) moist but slowly declining values during the winter, as the soil remains cool beneath the snowpack, largely isolated from any potential meltwater; followed by (ii) increasing wetness as snowmelt infiltrates into the soil, peaking at or just before the end of the melt period; and then (iii) steady declines in soil water content throughout the spring as evapotranspirative demand grows; leading to (iv) very dry conditions from late July through late September, especially at depth; followed by (v) a slow, steady wet-up period in the fall, with soil water contents peaking at most sites between late November and early December (Fig. 11). Soil water contents then slowly decline throughout winter, as precipitation is increasingly stored in the snowpack and soil water drains or is utilized in other ways. The persistent dryness in summer in the lower profile is notable, as it points to the lack of deep, summertime translocation events in these soils. In short, deep percolation events, which drive profile differentiation in podzolic soils, are

Table 3. Comparative data for the four National Weather Service sites along a strength of podzolization transect.†

Variable and/or Factor	Newberry	Gaylord	Houghton Lake	Mt. Pleasant
Strength of podzolization	Strong	Moderate-strong	Moderate-weak	Weak
Latitude	46°20′	45°01′	44°18′	43°36′
Mean maximum snow water equivalent (cm)	13.8	**16.2**	6.7	5.2
Mean number of contiguous days per winter with SWE > 0	**204**	189	171	170
Mean winter soil water content at 100 cm (m^3/m^3) (25 Dec–15 Feb)	**0.144**	0.147	0.149	0.155
Peak mean soil water content at 100 cm (m^3/m^3) (date)	**0.188** (2 Apr)	0.186 (8 Apr)	0.178 (16 Apr)	0.178 (6 Apr)
Mean difference in mean soil water content (wettest day minus driest day) (m^3/m^3)	**0.131**	0.130	0.122	0.123
Mean total wintertime (25 Dec–1 Mar) flux at 100 cm depth (cm); percent of mean annual flux in parentheses	**1.9 (7%)**	3.2 (9%)	3.0 (13%)	5.4 (17%)
Mean total snowmelt flux at 100 cm (15 Mar–10 May) (cm)	20.4 (61%)	**24.3 (61%)**	13.8 (55%)	14.9 (47%)
Mean total summertime flux at 100 cm (1 Jun–15 Sep) (cm)	0.9 (3%)	**0.7 (2%)**	0.8 (3%)	1.3 (4%)
Mean total autumn flux at 100 cm (15 Oct–5 Dec) (cm)	4.6 (13%)	**5.3 (12%)**	2.0 (7%)	3.1 (9%)

† Bolded numbers indicate the highest potential contribution to podzolization for this factor.

not common during the summer at any of the sites. Also notable is the fact that of the two times of the year when soils are wet (snowmelt and fall), soil water contents in the latter do not peak as high as during snowmelt (Fig. 11); snowmelt is the wettest time of year for these soils. These modeled data support the work of Schaetzl and Isard (1996), who argued that strong podzolization is at least partially associated with low soil water contents during the summer and strong inputs of water during snowmelt. As expected, soil water contents are much more dynamic in the upper profile than at depth (Fig. 11) because of small, frequent melt and rain events, many of which do not wet up the soil deeper than a few centimeters, but which may be effective at translocating soluble organic substances into the upper few centimeters of the mineral soil (see below).

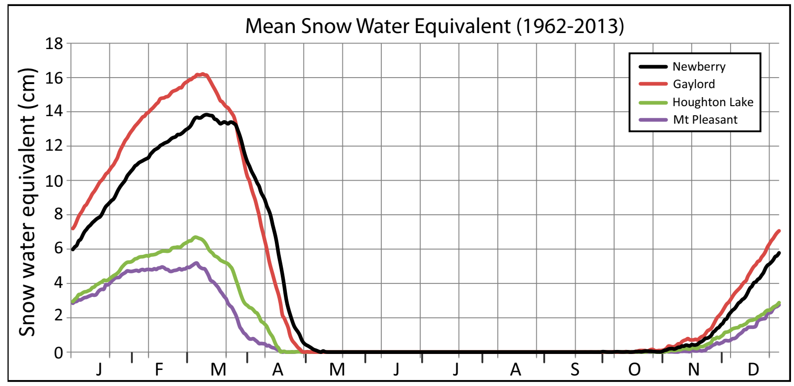

Fig. 10. Mean SWE data for the 1962–2013 period for the four NWS sites examined in this study.

Variations in soil water content along the transect can provide insight into its influence on podzolization. Soils at Newberry (the site associated with the strongest podzolization) are driest during the winter and wet up most rapidly to high levels during snowmelt (Fig. 11 and Table 3). Alternatively, the soils at Mt. Pleasant, the site furthest south, although still quite dry, are the wettest through the winter period and wet up much more slowly in spring, eventually attaining the lowest soil water contents of the four soils during the snowmelt period. The rapid increase in soil water content at Newberry, and to a lesser extent at Gaylord, may accentuate podzolization there. Dry soil conditions throughout winter may facilitate peeling and flaking of amorphous grain coatings in the upper profile, making them more likely to be flushed to the lower profile by rapid increases in soil water content associated with large meltwater fluxes during snowmelt. The simple act of wetting a dry soil has been linked to more efficient translocation of silicate clays (McKeague and St. Arnaud, 1969; Howitt and Pawluk, 1985). We suggest that the same logic works here: rapid wetting of a dry soil may facilitate the mobilization of both soluble substances and small fragments of organs mobilized in the upper profile, and their rapid translocation to the lower profile.

Equally interesting are soil water contents during the period of fall wet up (Fig. 11). The two sites that have the strongest podzolization (Gaylord and Newberry) remain consistently wettest throughout the fall. This finding is contrary to that reported by Schaetzl and Isard (1991), in which they observed that areas in the Lower Peninsula of Michigan with stronger podzolization had drier fall climates and soils. However, in a later paper, Schaetzl and Isard (1996) concluded only that shallow infiltration events in fall were typical of areas with strong podzolization, taking the emphasis off of soil water content and placing it on flux. We argue that wetter soils in fall can enhance podzolization in at least one key way: by keeping litter wet until a snowpack is established, thereby facilitating its decay and breakdown before the large snowmelt flux. Wetter conditions in fall may also facilitate small fluxes of water into the upper profile, carrying soluble organic materials released from the fresh litter. Newberry and Gaylord easily have the highest mean September to October precipitation totals (Fig. 3), leading to wetter soils (Fig. 11). Wetter soils here, at the onset of winter, coupled with deeper and more reliable snowpacks (Fig. 10), probably lead to slow, steady litter decomposition and a lower incidence of soil freezing, or only shallow freezing in years when it does occur. Wetter soils also have higher specific heat values, making freezing more difficult. Thus, the wetter soils in the north (in fall) have more potential

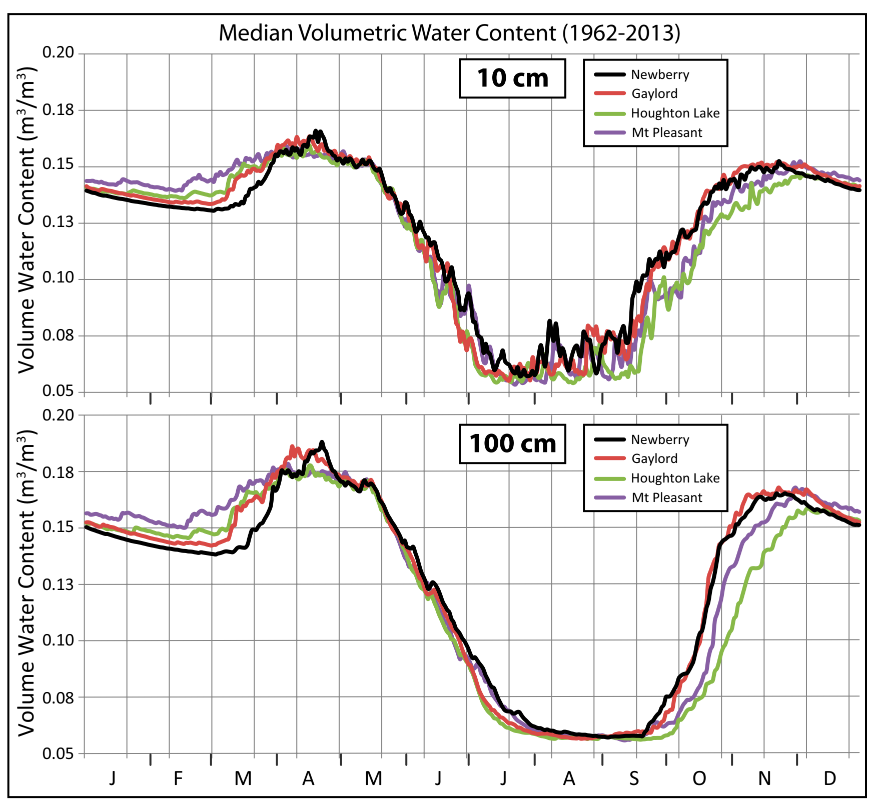

Fig. 11. Median contents of soil water (VWC) for the 1962–2013 period for the four NWS sites examined in this study.

to provide organic acids to infiltrating water during fall but most importantly during spring snowmelt, and this water has greater potential for deep percolation because of lowered incidences of soil freezing. Once again, the site with the weakest podzolization, Mt. Pleasant, has the driest soils in fall and, of course, the thinnest and most unreliable snowpacks.

Modeled flux data provide perhaps the most discriminating and useful information about podzolization across the transect (Fig. 12). Fluxes of water drive the translocation of soluble substances, the essence of podzolization. At all sites, shallow fluxes of water (into the uppermost profile [9 cm]) are common throughout the year, except for late December through mid-February when much of the precipitation is tied up in the snowpack (Fig. 12). At the more southerly sites, such fluxes in winter are considerably more common than in the north (as would be expected) because of warmer air temperatures which lead to brief episodes of snowmelt in midwinter. Regardless, water can and does enter the upper profile consistently throughout most of the year at all of the sites.

Most well-horizonated Spodosol profiles here have a B horizon that is at least 50 to 60 cm deep, and >30 cm thick. This type of morphology necessitates that wetting fronts penetrate to at least 100 cm; most of the well-developed Spodosols near Newberry have profiles at least that thick (Fig. 1 and Table 1). For this reason, we examined deep fluxes of water as a key factor in driving podzolization. We acknowledge that shallow flux events will keep the upper profile wet and enhance weathering there, but only by deep percolation can well-developed eluvial and illuvial zones be formed. Indeed, when deep fluxes of soil water (>100 cm) are considered, the distinctions along the transect become clear (Fig. 13).

Percolation fluxes to 100 cm clearly demonstrate the importance of the spring snowmelt period to podzolization. Across all four sites, at least 47%, and as high as 61%, of the total annual percolation flux occurred in the narrow window of time corresponding to spring snowmelt, with a clear decline from the two sites that have the strongest podzolization (Gaylord and Newberry) to the more weakly developed sites further south. Interestingly, the two sites where podzolization is strongest (Gaylord and Newberry) had not only the highest relative amount of snowmelt flux at 100 cm, but also had the highest absolute snowmelt flux totals: almost twice that of the two southerly sites (Table 3). In short, the two sites with strong podzolization were associated with a compression and partitioning of percolation fluxes into two discrete windows: a primary one during snowmelt and a secondary one in fall. Together, these two periods accounted for approximately three fourths of the total annual flux at Newberry and Gaylord (Fig. 13 and Table 3). In contrast, deep soil water fluxes at the more southerly

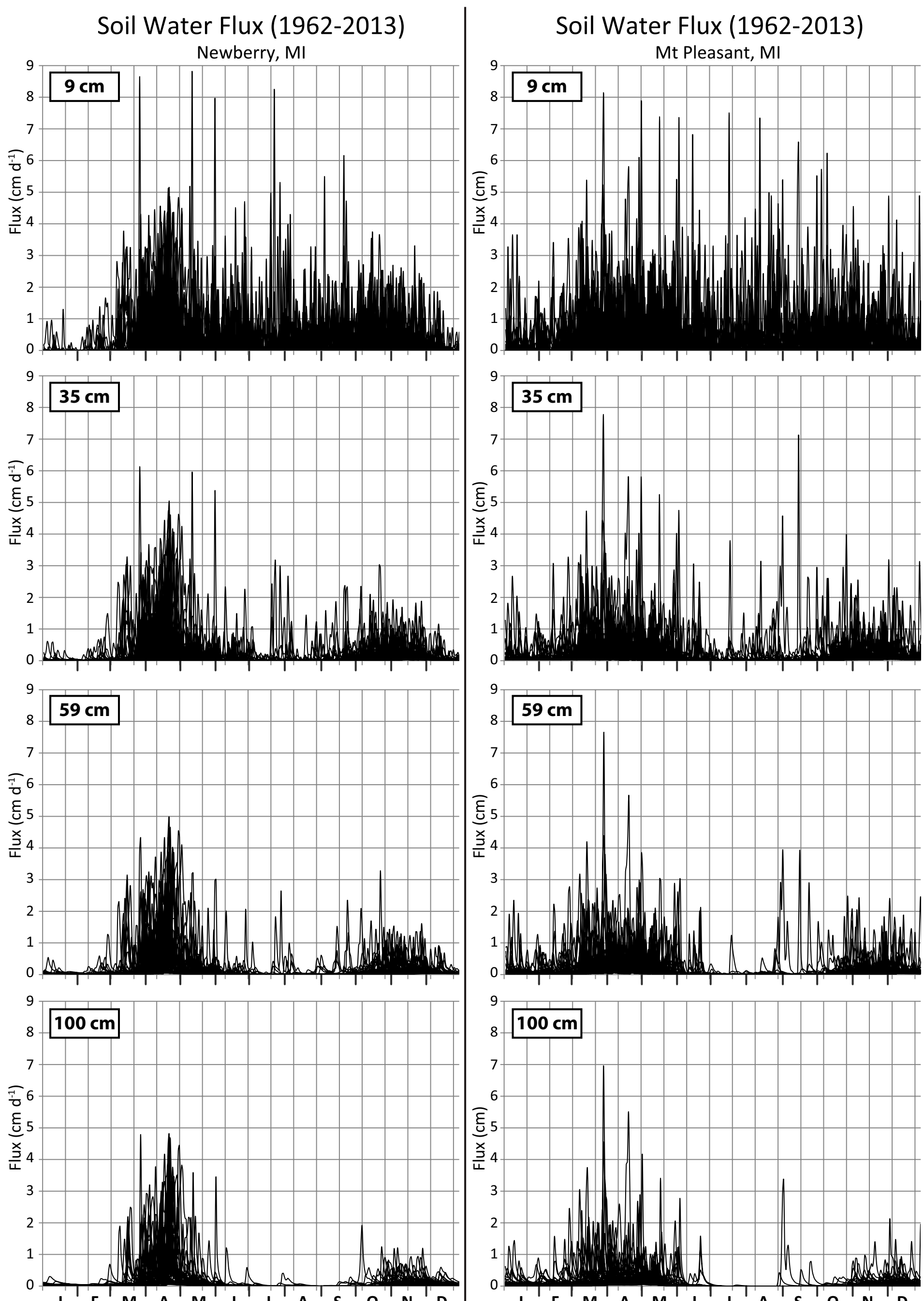

Fig. 12. Soil water flux data for the 1962–2013 period for Newberry (strongest podzolization) and Mt. Pleasant (weakest podzolization) for four soil depths. The data are exported from BROOK90 on a daily basis.

sites were distributed more evenly throughout the year, with the relative flux proportion for the snowmelt plus fall period amounting to only about 60% of the total annual flux. Slow, gradual flux throughout the winter was particularly noteworthy for Mt. Pleasant, accounting for 17% of the total annual flux (Fig. 13 and Table 3). Sites farther north had much smaller numbers of such flux events during winter.

In contrast to the deep and persistent snowmelt fluxes for these sites, our modeled data demonstrate that soils across the transect receive very few summertime flux events that penetrate to 100 cm (Fig. 11 and Table 3). Summertime fluxes of water to a depth of 100 cm are exceedingly rare, occurring only once and only three times for the 1961–2013 period at Mt. Pleasant and Newberry, respectively (Fig. 12). At most of the sites, an average of <1 cm of water (flux) reaches the 100-cm depth for the entire summer period (defined here as 1 June to 15 September), accounting for $\leq$4% of the total annual percolation flux across all of the sites (Table 3). As a result, the soils stay quite dry in summer, especially at depth, and podzolization (at least as it is driven by translocation) is paused (Fig. 11 and Table 3).

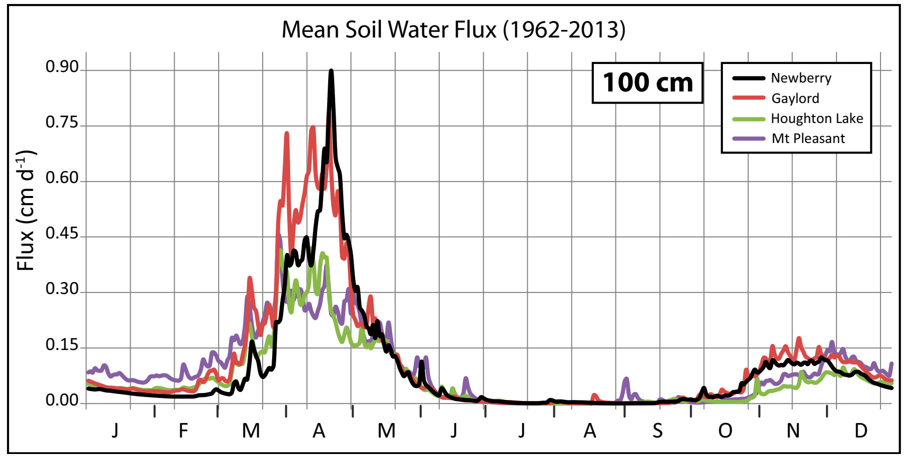

Fig. 13. Mean, modeled daily water flux data for the 1962–2013 period for the four NWS sites examined in this study.

Together, these data help explain why the best developed Spodosols in the Great Lakes region occur in association with snowbelts (Fig. 1). Areas within the snowbelts of the western Upper Peninsula, with weak Spodosols, are exceptions to the rule, mainly because the soils here are loamier. The strongest podzolization in the region occurs where snowbelts occur in association with sandy soils, as in the eastern Upper Peninsula, near our Newberry site, and to a lesser extent near Gaylord.

SUMMARY AND CONCLUSIONS

Daily NWS climate data for 1961–2013, input into a well-established hydrologic model, were used to estimate soil water contents and fluxes of water through soils along a climosequence of four stations in northern Michigan. This transect ranged from the area of strongest podzolization in the Great Lakes region, southward to the southern limit of observable podzolization. Modeled data were verified by comparing them to field data recovered from an instrumented location near the northernmost (Newberry) climate station.

Soils in this region maintain dry conditions during summer, especially in the lower profile. During summer, most wetting fronts (water fluxes) do not penetrate past the upper few centimeters of the mineral soil; cumulative summertime fluxes at 100 cm are usually <5% of the total annual flux at this depth. Thus, summer is a time of dry conditions and minimal translocation of substances associated with podzolization. Soils wet up slowly in fall, with the more northerly sites obtaining slightly wetter soil conditions. Nonetheless, deep fluxes of soil water are uncommon during the fall period. By midwinter, the soil wetness trend along the transect is reversed, as soils in the south receive small, episodic inputs of snowmelt, while the northerly sites dry down still further under a thick, cold snowpack. We suggest that the fall wet-up period is important to podzolization because the added water helps break down the fresh organic matter (litter) on the soil surface. Soluble organic materials released from the litter are then available for translocation into the mineral soil during the ensuing period of snowmelt.

The snowmelt period, with its large, often continuous, fluxes of cold meltwater, is a key component of podzolization in this region (Schaetzl and Isard, 1991, 1996). For most sites, and especially in the north, deep (>100 cm) fluxes of water are especially common during this period as snowpacks melt. Such fluxes comprise between 47 and 61% of the total annual flux and occur in only approximately 15% of the time. Variation in fluxes along the transect is predictable; sites in the north experience a later, more intense but shorter, snowmelt flux, with more overall water entering the soil and (importantly) penetrating to the lower profile. During spring, soils obtain their wettest conditions, and deep fluxes are most common. This type of snowmelt "pulse" drives podzolization, probably because (i) soluble organic materials from the O horizon are readily available for transport into the profile, with many having been released from the fresh litter in the previous fall, with more forming under the snowpack; (ii) only deep fluxes can differentiate the Spodosol profile by forming impoverished eluvial zones and deep B horizons enriched in organic materials and the metals brought with them; and (iii) the low temperatures of the snowmelt water inhibit microbial activity that could break down any organometallic complexes that are traveling in solution. Additionally, sites with large fluxes of snowmelt water have thicker snowpacks and, hence, less soil freezing, maintaining soil permeability during snowmelt. At southern sites, a larger proportion of snowmelt probably runs off and does not function directly in profile differentiation, that is, podzolization.

In sum, the long-held association between strong podzolization and thick snow cover appears to hold for sites in northern Michigan, which have some of the best developed Spodosols in the United States. Thick snowpacks insulate the soil and, when they melt, provide a strong and temporally compressed pulse of cold water, likely rich in soluble organics. This combination drives the podzolization process.

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