# Temperature of upland and peatland soils in a north central Minnesota forest

Dale S. Nichols

*USDA Forest Service, North Central Research Station, Grand Rapids, Minnesota 55744, USA.* *Received 9 April 1996, accepted 6 April 1998.*

Nichols, D. S. 1998. **Temperature of upland and peatland soils in a north central Minnesota forest**. Can. J. Soil Sci. **78**: 493–509. Soil temperature strongly influences physical, chemical, and biological activities in soil. However, soil temperature data for forest landscapes are scarce. For 6 yr, weekly soil temperatures were measured at two upland and four peatland sites in north central Minnesota. One upland site supported mature aspen forest, the other supported short grass. One peatland site was forested with black spruce, one supported tall willow and alder brush, and two had open vegetation — sedges and low shrubs. Mean annual air temperature averaged 3.6°C. Mean annual soil temperatures at 10- to 200-cm depths ranged from 5.5 to 7.6°C among the six sites. Soils with open vegetation, whether mineral or peat, averaged about 1°C warmer annually and from 2 to 3°C warmer during summer than the forested soils. The tall brush peatland was cooler than all other sites due to strong groundwater inputs. The mineral soils warmed more quickly in the spring, achieved higher temperatures in the summer, and cooled more quickly in the fall than the peat soils; however, the greatest temperature differences between mineral and peat soils occurred at or below 50 cm. In the upper 20 cm, vegetation and groundwater had greater effects on temperature than did soil type (mineral or peat). Summer soil temperatures were higher, relative to air temperature, during periods of greater precipitation. This effect was minimal at upland sites but substantial in the peatlands. In spite of the persistent sub-freezing air temperatures typical of Minnesota winters, significant frost developed in the soils only in those years when severe cold weather arrived before an insulating cover of snow had accumulated.

**Key words:** Soil temperature, vegetation effects, forest soils, groundwater, peatlands

Nicols, D. S. 1998. **Température des sols de hautes plaines et des sols de tourbières dans la forêt du centre-nord du Minnesota**. Can. J. Soil Sci. **78**: 493–509. La température du sol influe fortement sur les activités physiques, chimiques et biologiques du sol, mais on ne dispose malheureusement que de peu de données à cet égard sur les paysages forestiers. Durant six ans, nous avons mesuré les températures hebdomadaires du sol à deux emplacements en haute plaine et à quatre emplacements en tourbière daans le centre nord du Minnesota. L'un des emplacements en haute plaine portait un peuplement exploitable de trembles, l'autre une prairie à herbes courtes. Quant aux emplacements sous tourbière, l'un portait un peuplement d'épinettes noires, l'autre un embroussaillement haut de saules et d'aulnes et les deux derniers une végétation ouverte faite de carex et d'arbrisseaux bas. La température annuelle de l'air était de 3,6 °C, et la température annuelle moyenne du sol fluctuait de 5,5 à 7,6 °C selon l'emplacement. Les sols sous végétation ouverte, qu'il s'agisse de sol minéral ou de tourbe, étaient, en moyenne, de 1 °C par année et de 2 à 3 °C plus chauds en été que les sols forestiers. La tourbière à arbustes hauts était plus fraîche que tous les autres emplacements en raison du niveau élevé de la nappe. Les sols minéraux se réchauffaient plus rapidement au printemps et leurs températures estivales étaient plus hautes, mais ils se refroidissaient également plus vite en automne que les sols de tourbière. Toutefois, les différences de température les plus larges entre les deux types de sol s'observaient à la profondeur de 50 cm ou plus bas. Dans les 20 cm supérieurs du sol la végétation et hauteur de la nappe avaient plus d'effets sur la température que la nature du sol (minéral ou tourbe). La température estivale du sol était plus élevée par rapport à la température atmosphérique durant les épisodes de fortes précipitations. Cet effet était très faible aux emplacements de haute plaine mais plus prononcé dans les tourbières. Malgré les températures atmosphériques subgélives persistantes qui caractérisent les hivers au Minnesota, un gel significatif du sol ne s'observait que lorsqu'un hiver rigoureux survenait avant l'installation d'une bonne couche isolante de neige.

**Mots clés**: Température du sol, effet de la végétation, sol forestier, nappe, tourbière

Voluminous literature attests to the primary influence of temperature on physical, chemical, and biological activities in the soil, including water movement, evaporation, mineral weathering, organic matter accumulation and mineralization, nutrient cycling, seed germination, root growth and development, water and nutrient uptake, and many other processes. But despite the significance of soil temperature, such data are scarce, especially from forest land. For example, in the 1993 Annual Summary of Climatological Data for Minnesota (National Oceanic and Atmospheric Administration 1993), 154 stations reported air temperature and 189 reported precipitation amount, while only six reported soil temperature. All six soil temperature measurement stations had either bare ground or sod; none supported forest or other natural vegetation.

Peatland soil temperature data are particularly scarce, a conspicuous information gap, given that peatlands cover large areas of northern Europe, Asia, and North America. Minnesota, Wisconsin, and Michigan contain more than $6 \times 10^4$ $km^2$ ($1.5 \times 10^6$ acres) of peatland (Davis and Lucas 1959), which can comprise 20 to 40% of the land area in northern parts of these states. Farther north, over the

[1]*Mention of trade names does not constitute endorsement by the U.S. Department of Agriculture.*

357 000 $km^2$ of the Hudson Lowlands ecoregion, peatlands are by far the predominate ecosystem type, and an important type throughout the Taiga Shield and Boreal Shield Ecozones (Ecological Stratification Working Group 1995). Peatlands are becoming the focus of increasing attention as sites of environmentally significant processes such as carbon storage, methane production, and mercury cycling.

We measured soil temperatures over a period of 6 yr at six sites within or adjacent to the South Unit of the Marcell Experimental Forest. The Marcell Experimental Forest is located in north central Minnesota, about 25 miles north of Grand Rapids, on the Chippewa National Forest (Fig. 1). Two of the sites, one forested and one open (grass), are upland sites on well-drained mineral soils. The other four sites, one forested, one supporting tall brush, and two open (sedges and low shrubs), are in peatlands.

## MATERIALS AND METHODS

### Site Descriptions

UPLAND-ASPEN. This upland, forested, mineral soil site is in a well-stocked 70-yr-old aspen (*Populus tremuloides*, *Populus grandidentata*) stand with a site index of 20.7 m at 50 yr. As measured by Alban et al. (1994) on several plots 120–130 m south and east of this site, aspens make up about 88% of the aboveground biomass. Below the aspen canopy, the predominant shrub is beaked hazel (*Corylus cornuta*). The principal herbaceous plants are wild sarsaparilla (*Aralia nudicaulis*) and big-leaved aster (*Aster macrophyllus*). The soil at this site and at the adjacent Upland-Grass site is classified as a Cutaway loamy sand, a loamy, mixed Arenic Eutroboralf (Nyberg 1987). This soil consists of loamy sand to about 110 cm with clay loam material below that. A typical soil profile is shown in Fig. 2.

UPLAND-GRASS. This upland, open, mineral soil site is in an opening 60 m in diameter, created in 1978, in the aspen stand described above. The angle from ground level at the center of the opening to the tops of the surrounding trees is about 30 degrees. For most of the day, the site is not shaded by the surrounding aspen forest. The predominant vegetation is bluegrass (*Poa pratensis*) mixed with strawberry (*Fragaria virginiana*), aster (*Aster* spp.), wild pea (*Lathyrus ochroleucus*), and bracken fern (*Pteridium aquilinum*). The vegetation is mowed once or twice during the summer. The soil profile (Fig. 2) is the same as that described above for the Upland-Aspen site, except that the layer of forest-derived litter has been replaced by a thin layer of herbaceous litter.

S2 BOG. This peatland site is in a forested, ombrotrophic (nutrient-poor) bog. Incoming water and nutrients are derived mainly from direct precipitation. The bog is perched above the regional groundwater table and is not influenced hydrologically or chemically by groundwater. The overstory is black spruce (*Picea mariana*), about 125 yr old, that averages about 13 m in height and 15 cm dbh. The site index for black spruce is 9 m at 50 yr. Crown cover is about 60%. The understory consists of low, ericaceous shrubs — Labrador tea (*Ledum groenlandicum*) and leatherleaf (*Chamaedaphne calyculata*) and sphagnum moss — (*Sphagnum angustifolium*, *Sphagnum magellanicum*). Other common understory species include cotton grass (*Eriophorum spissum*), fine-leaved sedges (*Carex* spp.), and pitcher plant (*Sarracenia purpurea*) (Dise 1992). The soil at this site is classified as a Loxely peat, a Typic Borosaprist (Nyberg 1987). Because of the accumulation of organic acids and the low inputs of base cations and bicarbonates, the soil is very acidic. The average porewater pH is 3.9 at 0 to 40 cm (Dise 1992). The soil profile to 200 cm consists of mostly fibric and hemic materials to about 45 cm, underlain by highly decomposed sapric material (Fig. 2).

S3 FEN. This brushy peatland site is in a strongly minerotrophic (nutrient-rich) fen. This peatland intercepts the regional groundwater table, which supplies significant amounts of water as well as mineral nutrients to the fen. These groundwater-borne minerals, principally base-cations and bicarbonate, result in a higher porewater pH (about 6.0–6.5) (Verry 1975) and greater vegetative productivity than at the other peatland sites. Before it was harvested in 1973, this site had an overstory of 105-yr-old black spruce (*Picea mariana*), with smaller amounts of northern white cedar (*Thuja occidentalis*), balsam fir (*Abies balsamea*), and tamarack (*Larix laricina*) (Knighton and Steigler 1981). The fen was clearcut in January 1973, and the slash was broadcast and burned in July 1973. In the winter of 1974, the area was seeded with black spruce. The present plant community is dominated by a dense shrub layer of speckled alder (*Alnus rugosa*) and willow (*Salix* spp.), about 3 to 4 m in height. Below the shrubs are some 10 000 stems per hectare of black spruce, about 2 m high. The soil is classified as a Lupton peat, a Typic Borosaprist (Nyberg 1987). The soil profile to 200 cm consists almost entirely of sapric material (Fig. 2). Existing fibers are primarily woody.

BOG LAKE FEN. This peatland site is an open, weakly minerotrophic fen. This peatland receives some surface and subsurface flow from the surrounding upland. The average porewater pH of 4.6 (Clement et al. 1995) is slightly higher than that at the S2 Bog site and reflects the influence of small inputs of base cations in runoff and/or groundwater. This peatland is wetter (has a higher water table) than the S2 Bog and S3 Fen sites. The higher water table is responsible for the lack of tree cover on this fen. Sphagnum moss (predominately *Sphagnum papillosum*) covers about 60% of the peatland surface, especially in the hollows. Vascular plants are largely confined to the hummock portions of the hummock/hollow topography. The principal vascular plants are arrowgrass (*Scheuchzeria palustris*), beak-rush (*Rhynchospora alba*), and pitcher plant (*Sarracenia purpurea*). Other common species include sedge (*Carex paupercula* and *Carex oligosperma*), leatherleaf (*Chamaedaphne calyculata*), and bog laurel (*Kalmia polifolia*) (Clement et al. 1995). The soil is mapped as a Greenwood peat (Nyberg 1987). Greenwood peat is classified as a Typic Borosaprist; however, the soils at this site contain significantly more fibrous material than the soils at the S2 and S3 sites. A typical soil profile to 200 cm consists

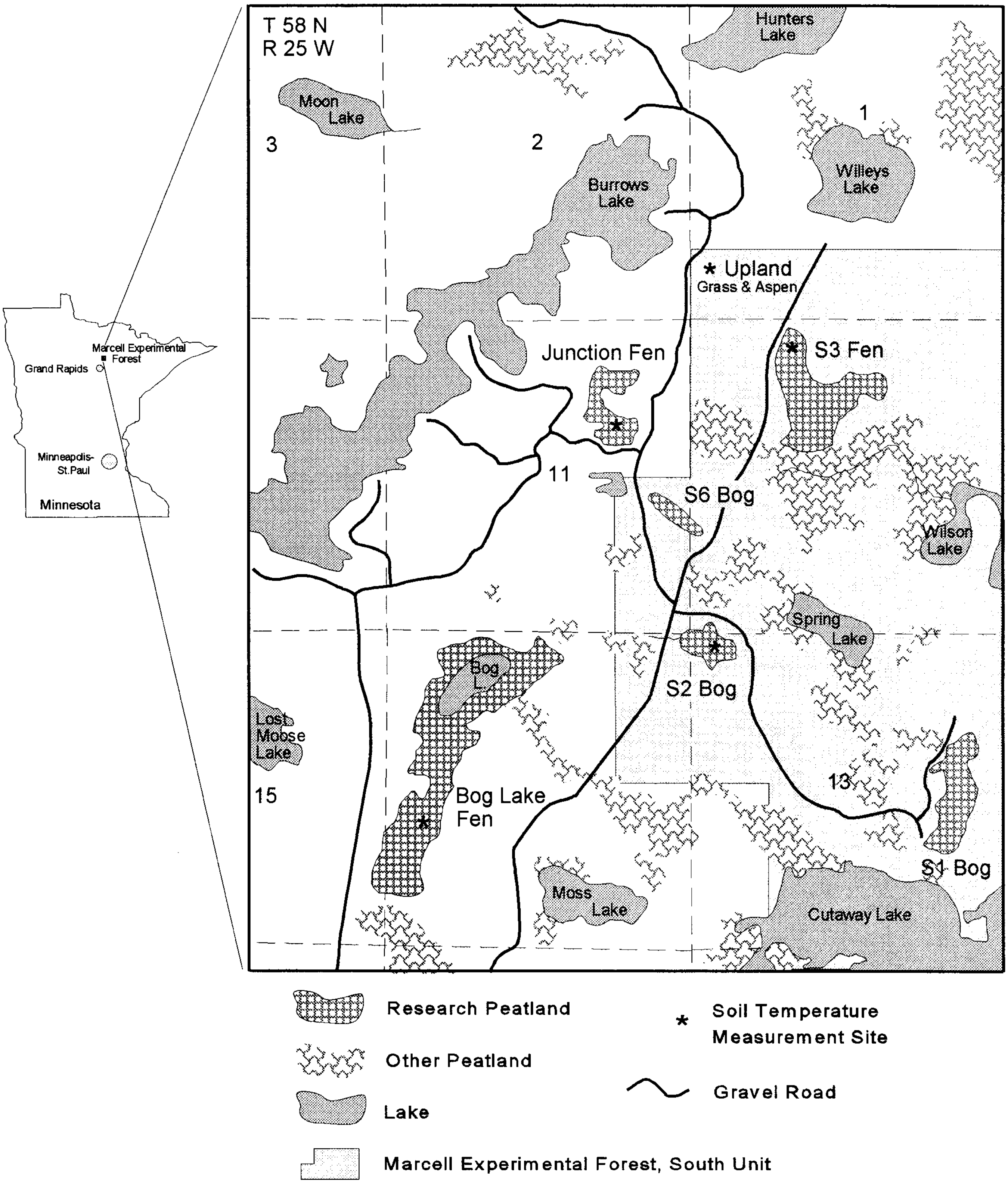


**Fig. 1**. Map of the Marcell Experimental Forest showing locations of soil temperature measurement sites.

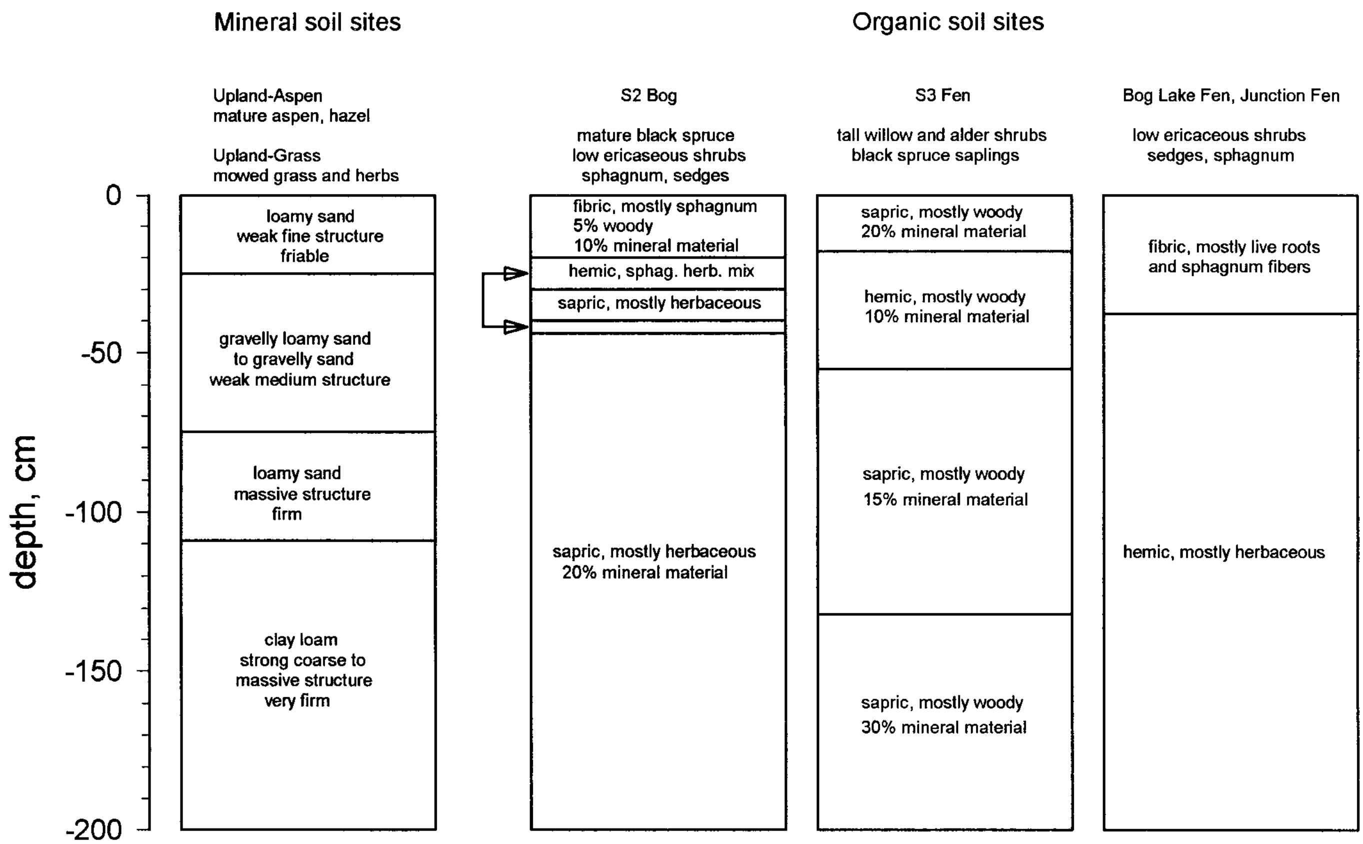


**Fig. 2**. Soil profiles of study sites on the Marcell Experimental Forest.

of fibric material to about 40 cm, underlain by hemic material (Fig. 2). Fibers are mostly herbaceous.

JUNCTION FEN. This peatland site is an open, weakly minerotrophic fen, similar to Bog Lake Fen. The primary vascular plant species are sedge (*Carex oligosperma*), arrowgrass (*Scheuchzeria palustris*), and cranberry (*Vaccinium oxycoccus*) over a ground cover of sphagnum moss (*Sphagnum angustifolium*, *Sphagnum capillifolium*, and *Sphagnum fuscum*) (Dise 1992). Like Bog Lake Fen, this peatland receives small amounts of mineral nutrients from upland runoff and/or groundwater. The porewater pH averages about 4.5 (Dise 1992). Also like Bog Lake Fen, Junction Fen is too wet to support a forest cover. The soil at this site has not been classified to the series level, but a typical soil profile is similar to the one described above for the Bog Lake Fen site.

## Soil Temperature Measurements

Soil temperatures were measured using thermocouples constructed from insulated type T (copper/constantan) wire with a ±0.5°C limit of error. At each of the six sites, thermocouples were installed in the soil at 10, 20, 30, 40, 50, 100, and 200 cm below the surface. The four peatlands studied have a hummock/hollow microtopography with a relief of 10 to 30 cm between the tops of the hummocks and the bottoms of the hollows. At all four peatland sites, the elevations (above mean sea level) of the bottoms of the hollows are much more uniform than that of the tops of the hummocks. In this paper, the term peatland “surface” refers to the mean elevation of the bottom of the hollows. We facilitated installation by inserting individual thermocouples through holes drilled in a 1.6-cm-diameter, 2-meter-long, wooden dowel at the appropriate spacing. The dowel with the stack of thermocouples was then inserted into the soil. In the peat soils, we were able to accomplish this with almost no disturbance by first driving a metal rod of a similar diameter into the soil, which, when removed created a pilot hole into which the stack of thermocouples could be easily inserted. In the mineral soils, the thermocouple stacks were placed into holes that were augered into the ground and then backfilled.

Soil temperatures were measured weekly at each site, usually between 10:00 h and 13:30 h. At all but the Upland-Aspen site, an Omega hand-held model HH-25TC digital type-T thermocouple thermometer was used to read the thermocouples. At the Upland-Aspen site, where electricity is available, the thermocouples were permanently attached to an Omega model 650 bench top, 10-channel, digital type-T thermometer.

The accuracy of both thermocouple thermometers was tested over a range of ambient temperatures. The meters were allowed to equilibrate with air temperatures of from –20 to +40°C, and temperature measurements were recorded from thermocouples immersed in deionized ice water having a temperature of 0°C. Instrument error for the HH-25TC ranged from about –0.8°C at an ambient (meter) temperature of –20°C, to about +0.4°C at +40°C. Regardless of ambient temperature, the model 650 bench top instrument consistently indicated a temperature of 32°F (0°C) for deionized ice water (this instrument displays thermocouple temperature only to the nearest degree F). When the thermocouple was exposed to air temperatures over the range of –20 to +40°C, the temperature of the thermocouple made no measurable difference in the accuracy of either meter.

When soil temperatures were measeured in the field, the accuracy of the meter was checked with deionized ice water at each site. Where necessary, winter soil temperatures collected during bitter cold weather were corrected for meter error. On cold days, the most accurate readings were obtained when the meter was carried outside of the clothing and allowed to remain in equilibrium with the air temperature. Additional error was introduced if the meter was removed from a warm pocket before use and the temperature of the meter was changing while measurements were being taken.

At the Upland-Aspen site, the Omega model 650 meter used was mounted inside a small insulated shelter heated by a light bulb during cold weather. Except on two or three occasions when the light bulb burned out, the temperature in the instrument shelter did not fall below 10°C. On these occasions, it is not likely that the thermocouple temperature readings were affected.

### Other Measurements

Air temperature on the South Unit of the Marcell Experimental Forest has been measured continuously since 1961 by a recording hygrothermograph housed in a standard United States National Weather Service shelter. The hygrothermograph is calibrated weekly against a mercury maximum-minimum thermometer. The shelter is located in an opening in the aspen forest, about 200 m northeast of the S2 Bog site (Fig. 1). Mean daily air temperatures are calculated as means of the minimum and maximum temperatures occurring in each 24-h period from midnight to midnight.

A US National Weather Service standard 20.3 cm (8.0 in) diameter rain gauge is also maintained at this site . Fitted into the gauge orifice is a 20.3 cm funnel that concentrates the rain into a 5.1 cm (2.0 inch) diameter receiver, allowing measurements to 0.025 cm (0.01 in). For snow collection, the funnel and receiver are removed.

Groundwater elevation is measured at several sites on the Marcell Experimental Forest, including a continuously measured site designated as well 305, located on the upland adjacent to S3 Fen, about 120 m northwest of the S3 Fen temperature measurement site (Fig. 1). The water table is about 6 m below the ground surface at well 305. Water level is also continuously measured in wells in S3 Fen, S2 Bog, and Bog Lake Fen, (but not in Junction Fen) located near the respective temperature measurement sites.

## RESULTS AND DISCUSSION

### Air Temperature

Northern Minnesota has a marked continental climate characterized by short, warm summers; long, cold winters; and wide variations in temperature. Annual maximum temperatures are typically in the 30 to 35°C range, while the minimum for the year is usually around –35 to –40°C. From 1961 to 1995, mean annual air temperature at the Marcell Experimental Forest ranged from 1.5 to 6.1°C, with an average of 3.1°C. For the 6-yr period, 1990–1995, during which soil temperatures were measured, mean annual air temperature ranged from 2.5 to 4.3°C, and averaged 3.5°C. In comparison to the period of record, air temperatures in 1990–1995, were a little warmer in the winter and spring and cooler in the fall than normal. December through June averaged about 1.0°C warmer, and October and November about 0.8°C cooler than the 35-yr mean for these months. For the July through September period, the 1990–1995 mean was within 0.1°C of the 35-yr mean.

### Annual Soil Temperature Cycle

Figures 3 and 4 show the annual cycle of soil temperature at the six sites. The data shown are semi-monthly averages (each month is divided into two periods, days 1 through 15, and days 16 through the end of the month) of soil temperature at each depth and site, and air temperature, over the 6-yr period. The data are grouped by site in Fig. 3 and by depth in Fig. 4. The 6 yr of soil temperature data are also summarized in Table 1, which shows mean summer (June–August), winter (December–February), and annual temperatures at five depths (10, 20, 50, 100, 200 cm) at each of the six sites.

The coldest period of the year is between the last week of December and the first week of February, during which time the air temperature averages about –15°C. Under the cover of snow, soil temperatures at the six sites fluctuated relatively little during the winter months. From January through March, soil temperatures at 10 cm and 20 cm below the surface hovered near 0°C. The development of frost varied from year to year, depending on whether an insulating cover of snow accumulated before the onset of cold winter air temperatures. Temperatures at 50, 100, and 200 cm fell slowly during the winter, but remained warmer than near-surface temperatures. Soil temperatures from January through March averaged 0.9, 1.7, and 3.2°C at 50, 100, and 200 cm below the surface of the mineral soils and 2.2, 4.1, and 6.3°C at the same depths in the peat soils.

Snow depth and frost development in the soil are measured in late February each year at the Marcell Experimental Forest at a total of 100 sites in three cover types — open (which includes sites in Junction bog), conifer forest (which includes sites in S2 Bog), and deciduous forest (sites similar to the Upland-Aspen site in the present study). Monthly winter air temperature, snow depth, and frost data for 1990–1995 are presented in Table 2. Frost developed to a depth of at least 10 cm in 3 (1990, 1991, 1995) of the 6 yr of this study and exceeded 20 cm in 1990. In these years, cold winter temperatures occurred before an adequate insulating snow cover had accumulated. In the winter of

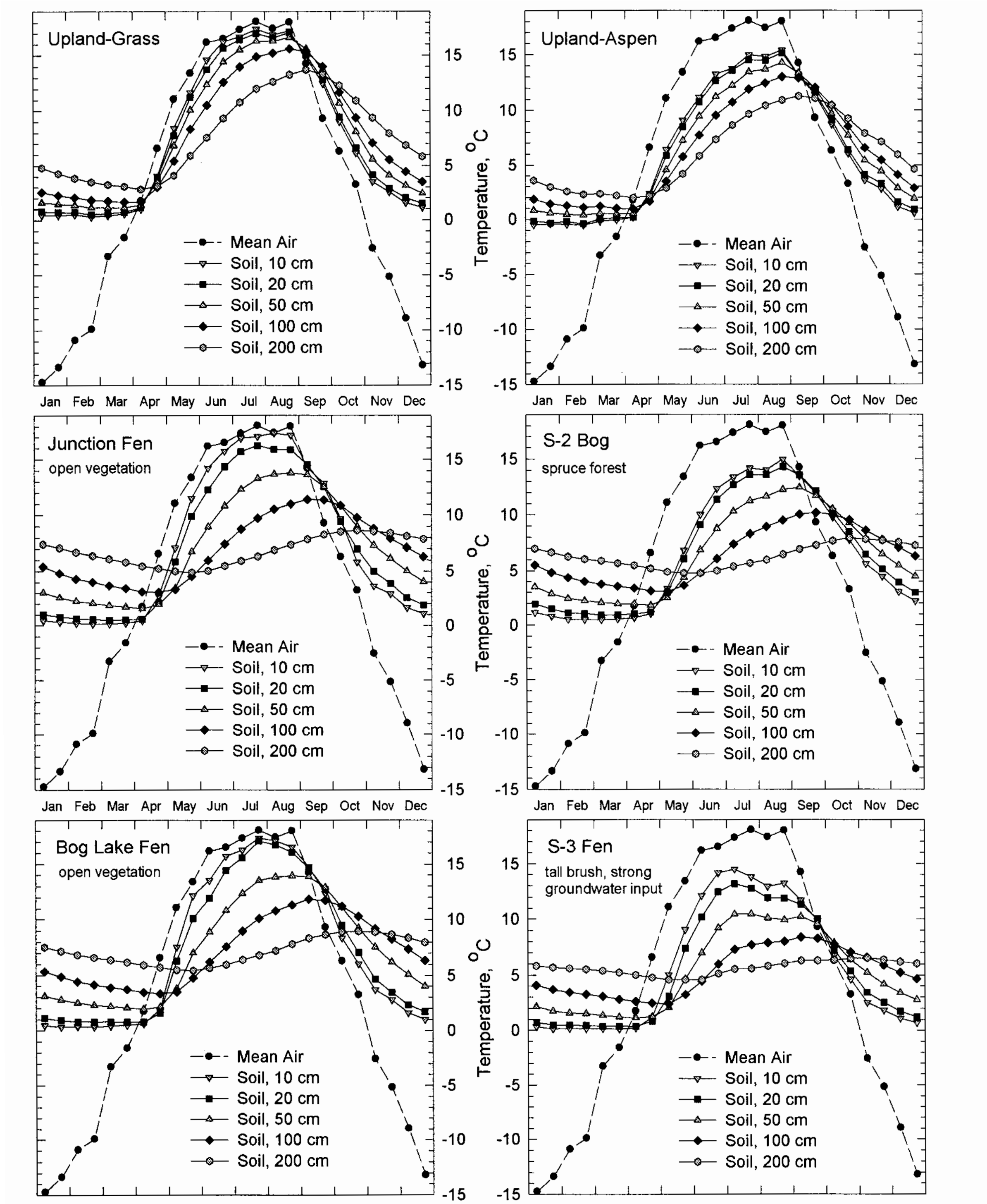


**Fig. 3**. Average daytime soil temperatures, Marcell Experimental Forest, 1990–1994, grouped by site.

1989–1990, particularly cold weather came in early December. In 1992, 1993, and 1994, heavier snows early in the winter resulted in only shallow and incomplete frost. Verry (1991) evaluated 30 yr of frost records from several upland and peatland sites on the Marcell Experimental Forest and concluded that if the snow pack is more than 20 cm deep in late December, frost will penetrate no more than 20 cm into the soil and usually less. Late December snowpacks of 10 cm or less, in conjunction with bitter cold, can produce frost in excess of 40 cm.

By the middle of February, average air temperature begins to rise steadily. The mean daily air temperature

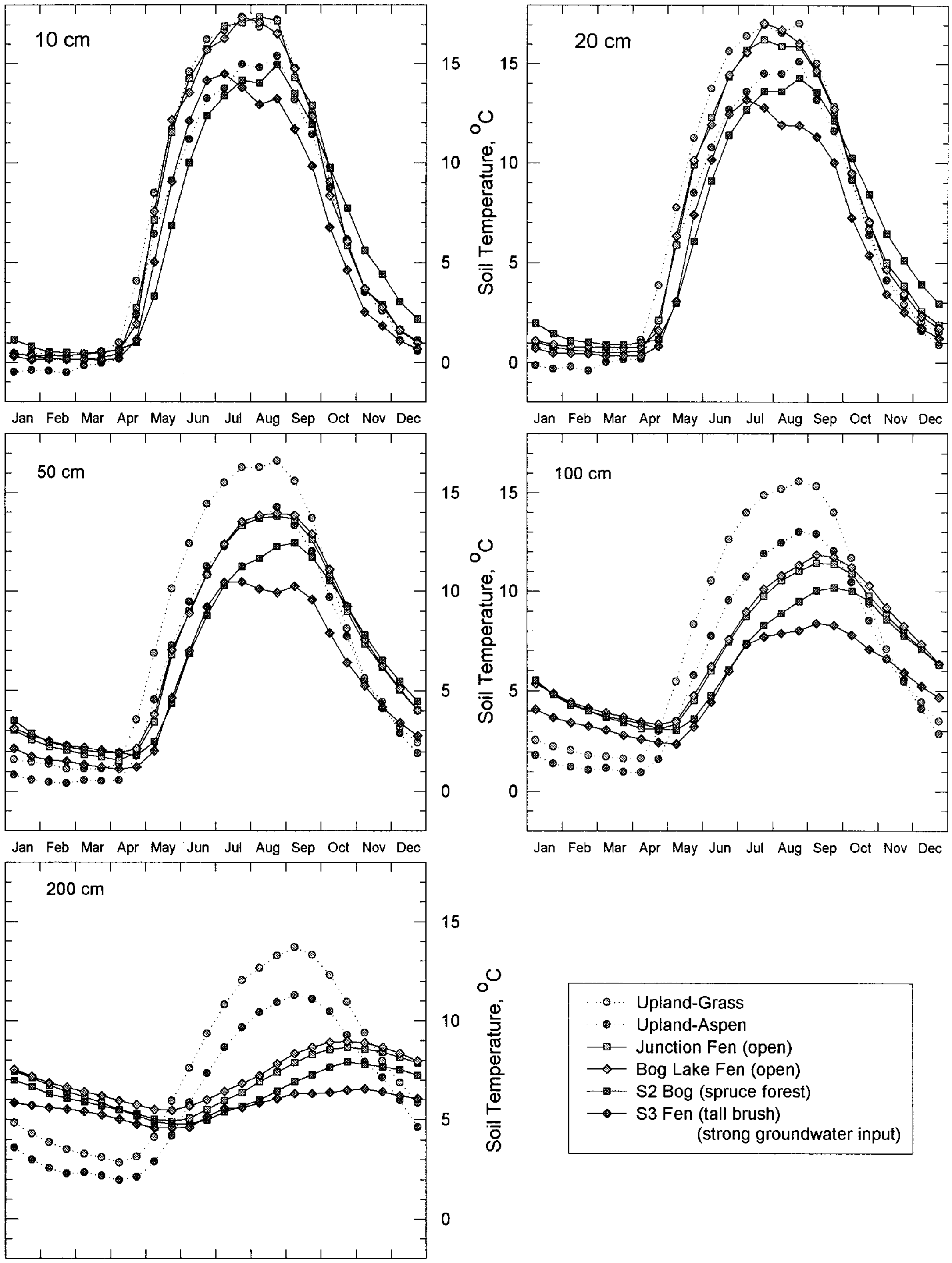


**Fig. 4**. Average daytime soil temperatures, Marcell Experimental Forest, 1990–1994, grouped by depth.

exceeds 0°C by early April, on the average. Most years the snow pack is gone by late March to mid-April, as is any frost that developed in the soil over the winter. In some years, the occurrence of cold weather after the insulating layer of snow has melted will cause frost to reform as late as early May, especially in the peatlands (Verry 1991). As spring progresses, air temperature continues to increase in an almost linear fashion until well into June. The upland soils responded quickly. By mid-April they were already beginning to warm. At the grassed upland site, soil temperatures down to 50 cm increased almost as rapidly as mean daily air temperature. Temperatures at 10 and 20 cm in the

**Table 1. Mean annual, summer, and winter temperatures of two mineral and four organic soils, Marcell Experimental Forest, Minnesota, 1990–1995**

| (Site) | Upland-Open | Junction Fen | Bog Lake Fen | Upland-Aspen | S2 Bog | S3 Fen |
|---|---|---|---|---|---|---|
| (Soil) | Mineral | Peat | Peat | Mineral | Peat | Peat |
| (Cover) | Grass | Sedge/shrub | Sedge/shrub | Aspen | Spruce | Tall brush |
| | | | *Mean annual temperature ± 95% CI (°C)* | | | |
| Depth | | | | | | |
| 10 cm | 7.47 ± 0.77 | 7.29 ± 0.83 | 7.05 ± 0.82 | 6.21 ± 0.70 | 6.32 ± 0.62 | 5.60 ± 0.65 |
| 20 cm | 7.49 ± 0.74 | 7.12 ± 0.69 | 7.07 ± 0.77 | 6.20 ± 0.68 | 6.43 ± 0.57 | 5.35 ± 0.58 |
| 50 cm | 7.68 ± 0.68 | 7.06 ± 0.52 | 7.11 ± 0.56 | 6.27 ± 0.60 | 6.38 ± 0.44 | 5.16 ± 0.42 |
| 100 cm | 7.65 ± 0.60 | 6.98 ± 0.33 | 7.09 ± 0.37 | 6.20 ± 0.52 | 6.32 ± 0.29 | 5.18 ± 0.25 |
| 200 cm | 7.66 ± 0.43 | 6.82 ± 0.14 | 7.06 ± 0.15 | 6.19 ± 0.39 | 6.22 ± 0.12 | 5.60 ± 0.08 |
| Mean | 7.59 ± 0.29 | 7.13 ± 0.26 | 7.12 ± 0.26 | 6.33 ± 0.25 | 6.21 ± 0.20 | 5.38 ± 0.20 |
| Tukey HSD[z] all depths | *a* | *a* | *a* | *b* | *b* | *c* |
| | | | *Mean summer (June-August) temperature ± 95% CI (°C)* | | | |
| Depth | | | | | | |
| 10 cm | 16.51 ± 0.35 | 16.08 ± 0.49 | 15.58 ± 0.57 | 14.08 ± 0.46 | 12.98 ± 0.46 | 13.27 ± 0.41 |
| 20 cm | 16.03 ± 0.37 | 14.90 ± 0.42 | 14.87 ± 0.61 | 13.66 ± 0.45 | 12.24 ± 0.48 | 11.96 ± 0.38 |
| 50 cm | 15.12 ± 0.40 | 11.99 ± 0.45 | 11.78 ± 0.59 | 12.42 ± 0.44 | 9.98 ± 0.48 | 9.36 ± 0.37 |
| 100 cm | 13.63 ± 0.45 | 8.81 ± 0.46 | 8.83 ± 0.50 | 10.89 ± 0.46 | 7.28 ± 0.41 | 6.73 ± 0.33 |
| 200 cm | 10.78 ± 0.48 | 6.13 ± 0.18 | 6.49 ± 0.21 | 8.78 ± 0.44 | 5.42 ± 0.15 | 5.37 ± 0.13 |
| Tukey HSD 20 cm | *a* | *b* | *b* | *c* | *d* | *d* |
| | | | *Mean winter (Dec.-Feb.) temperature ± 95% CI (°C)* | | | |
| Depth | | | | | | |
| 10 cm | 0.56 ± 0.14 | 0.65 ± 0.15 | 0.72 ± 0.18 | -0.04 ± 0.24 | 1.07 ± 0.23 | 0.35 ± 0.14 |
| 20 cm | 1.02 ± 0.15 | 1.28 ± 0.19 | 1.29 ± 0.18 | 0.24 ± 0.26 | 1.96 ± 0.62 | 0.77 ± 0.16 |
| 50 cm | 1.84 ± 0.19 | 3.19 ± 0.25 | 3.36 ± 0.26 | 1.21 ± 0.26 | 3.40 ± 0.62 | 2.13 ± 0.19 |
| 100 cm | 2.74 ± 0.23 | 5.36 ± 0.26 | 5.49 ± 0.28 | 2.14 ± 0.28 | 5.26 ± 0.47 | 3.99 ± 0.18 |
| 200 cm | 4.85 ± 0.27 | 7.32 ± 0.15 | 7.41 ± 0.30 | 3.75 ± 0.31 | 6.80 ± 0.32 | 5.78 ± 0.08 |

*a–d* Letters below columns indicate sites that are not significantly different from each other at the 5% level.

soil lagged only a week or two behind mean daily air temperature. The soil at the forested upland site began to warm almost as early as at the adjacent open site, but remained 2 to 3°C cooler than at the grassed site at all depths. At both upland sites, heat moved quickly down through the mineral soil profiles, and by the end of April, soil temperatures down to 200 cm were increasing. The lag time between the onset of springtime warming at 10 cm and at 200 cm was only about 2 wk.

As can be seen in Figs. 3 and 4, the peatland soils warmed more slowly than the upland soils. Down to 20 cm, peatland soil temperatures were only a week or two behind upland temperatures, and eventual summer soil temperatures were similar, especially in the open sites. But at 50 cm and below, peatland soils lagged farther behind the uplands. At 50 cm in the peat soils, there was little or no warming until early May, and at 100 cm, temperatures did not start to rise until toward the end of May. At 200 cm, temperatures continued to decrease until late May or even early June before beginning to rise again, about 2 mo after the surface layers began to warm.

Summers in northern Minnesota are short. From the last part of June through the last part of August, air temperatures average about 18°C, the warmest days of the year typically occurring in mid-July. At 10 cm below the surface, the mean summer (June–August) soil temperature at the Upland-Grass site was 16.5°C, compared with an average of 15.8°C for the two open peatland sites — Junction Fen and Bog Lake Fen (Table 1). For the forested sites, the mineral soil (Upland-Aspen) averaged 14.1°C compared with 13.0°C at 10 cm in the peat soil (S-2 Bog). Deeper in the soil, the summer temperature differences between mineral and peat soils were greater. At 50 cm, summer temperature averaged 15.1°C at the grassed upland site compared with 11.9°C at the open peatland sites, and 12.4°C compared with 10.0°C at the forested upland and peatland sites, respectively. At 100 cm, the grassed upland site averaged 13.6°C compared with 8.8°C in the open peatlands, and 10.9°C in the upland aspen forest compared with 7.3°C under the spruce forest canopy in S2 Bog.

By early September, mean daily air temperatures are declining rapidly. Soil cooling patterns in autumn were similar to spring warming patterns. The mineral soils cooled quickly, with temperatures down to 200 cm beginning to decrease by the last half of September. The surface layers of the peat soils cooled rapidly as well, but deeper in these soils, the transition from warming to cooling was delayed. At 200 cm in the peatland sites, the warmest temperatures of the year occurred in late October or early November.

The rapid autumnal decline of near-surface soil temperatures continued until early November when, on the average, snow cover began to accumulate and insulate the soil, slow-

**Table 2. Snow cover and frost depth at the Marcell Experimental Forest, 1990–1995**

| | | Mean air temp. (°C) | | | Snow depth (cm) | | | | | | | | Frost on 23Feb.[z] | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Winter | Cover type | Dec | Feb | Mar | 1 Dec. | 15 Dec. | 1 Jan. | 15 Jan. | 1 Feb. | 15 Feb. | 1 Mar. | 23 Feb.[z] | Depth (cm) | Occurrence (%) |
| | Lawn[y] | | | | 10 | 13 | 23 | 23 | 41 | 33 | 41 | | | |
| 1989–1990 | Open | –23.5 | –15.1 | –20.5 | | | | | | | | 42 | 25 | 100 |
| | Deciduous | | | | | | | | | | | 40 | 23 | 100 |
| | Conifer | | | | | | | | | | | 36 | 23 | 100 |
| | Lawn | | | | 3 | 3 | 20 | 28 | 30 | 28 | 36 | | | |
| 1990–1991 | Open | –19.8 | –22.6 | –15.5 | | | | | | | | 40 | 13 | 95 |
| | Deciduous | | | | | | | | | | | 41 | 15 | 96 |
| | Conifer | | | | | | | | | | | 33 | 15 | 100 |
| | Lawn | | | | 8 | 20 | 20 | 25 | 25 | 28 | 28 | | | |
| 1991–1992 | Open | –16.5 | –15.1 | –12.0 | | | | | | | | 48 | 5 | 50 |
| | Deciduous | | | | | | | | | | | 44 | 8 | 58 |
| | Conifer | | | | | | | | | | | 39 | 5 | 50 |
| | Lawn | | | | 3 | 15 | 33 | 41 | 38 | 36 | 38 | | | |
| 1992–1993 | Open | –17.5 | –20.8 | –18.4 | | | | | | | | 50 | 3 | 55 |
| | Deciduous | | | | | | | | | | | 42 | 5 | 58 |
| | Conifer | | | | | | | | | | | 37 | 5 | 77 |
| | Lawn | | | | 23 | 23 | 38 | 46 | 48 | 53 | 28 | | | |
| 1993–1994 | Open | –15.9 | –28.0 | –21.9 | | | | | | | | 57 | 5 | 75 |
| | Deciduous | | | | | | | | | | | 54 | 8 | 78 |
| | Conifer | | | | | | | | | | | 47 | 5 | 70 |
| | Lawn | | | | 10 | 15 | 13 | 23 | 30 | 48 | 38 | | | |
| 1994–1995 | Open | –13.6 | –18.1 | –20.8 | | | | | | | | 52 | 10 | 85 |
| | Deciduous | | | | | | | | | | | 48 | 10 | 100 |
| | Conifer | | | | | | | | | | | 40 | 10 | 85 |

[z]Average date, varied from 16 February to 1 March.
[y]Weather station at Grand Rapids, MN, 40 km south of the Marcell Experimental Forest.

ing further heat loss. Once a sufficient depth of snow accumulated, winter air temperatures, which commonly fall to –40°C or below, have relatively little effect on soil temperature.

Mean annual soil temperature, measured at 10 to 200 cm below the surface, ranged from 7.6 to 5.5°C (average of the five depths) among the six sites (Table 1). Because more solar energy reaches the ground there, the soils at the open-vegetation sites were warmer than at the forested sites. These warmer temperatures were seen at all depths measured and in both mineral and organic soils. The open sites (Upland-Grass, Junction Fen, Bog Lake Fen) averaged about 1°C warmer than the forested sites (Upland-Aspen, S-2 Bog) on an annual basis. The grassed mineral soil site was also almost 0.5°C warmer than the open peatland sites, but this difference was not statistically significant. The two forested sites (aspen on mineral soil and black spruce on peat) were not significantly different from one another. S3 Fen (tall brush peatland) was significantly cooler than all the other sites due to the strong input of groundwater at this site. Differences between the open sites and forested sites, and between S3 Fen and the other sites, were statistically significant at the 5% level, based on a Tukey pairwise comparison of means (Steel and Torrie 1960). Since mean annual soil temperature did not vary significantly with depth at any site, the mean annual temperatures for all depths at each site were lumped together for these comparisons.

In the summer, the open sites were generally from 2 to 3°C warmer than the forested sites. A Tukey comparison of temperatures at the 20-cm depth indicates that the summer temperature difference between forested and open sites was significant. The temperatures differences between mineral and peat soils with open vegetation, about 1°C at 20 cm, were also significant, as were those between the mineral and peat soil forested sites. Near the surface, during the summer, S3 Fen was significantly cooler than all the other sites except S2 Bog.

While soil temperatures at the upland aspen forest site and the S2 Bog spruce forest site were nearly the same, on a mean annual basis, as were the soil temperatures at the open upland and open peatland sites, the total range in temperature over the year was greater in the mineral soils than in the peat soils, and the mineral soils warmed more quickly in the spring, achieved higher temperatures in the summer, cooled more rapidly in the autumn, and were slightly colder in the winter than the peats (Figs. 3 and 4). These differences were minimal near the surface, but increased significantly with depth.

It is well known that at a given rate of heat input, the change in temperature exhibited by a soil and the rate at which that change occurs depend upon the heat capacity and thermal conductivity of the soil. Baver et al. (1972) present a good discussion of this topic and cite references dating back as far as 1878. Heat capacity is the amount of heat, in calories, needed to raise 1 $cm^3$ of soil 1°C, and is approximately equal to $0.46x_m + 0.60x_o + 1.0x_w$, where $x_m$, $x_o$, and $x_w$ are the volume fractions of mineral material, organic matter, and water in the soil. Thermal conductivity is the rate at which heat moves through the soil. On the average, thermal conductivity of different soils is in the order of sand > loam > clay > peat. Thermal conductivity is higher in soils with larger particles, and with higher bulk densities, and is higher in moist soils than in dry. The thermal conductivity of wet peat is only about 60% of that of moist mineral soil.

Given that water and peat require more heat to raise their temperature than does an equal volume of mineral soil, and that peat conducts heat more slowly than does mineral soil, I generally anticipated the basic thermal differences between mineral and organic soils demonstrated by the present study, i.e., that organic soils warm and cool more slowly, fluctuate over a more narrow temperature range, and are colder during the growing season than mineral soils. The magnitude of these differences, however, is worth some discussion.

Because of high water tables, the upper 20 or 30 cm of the peat, plus the hummocks, constitutes the bulk of the vascular plant rooting zone in these peatlands. The study showed that down to about 20 cm, summer peat temperatures were not much cooler (only about a 1°C difference) than those in the mineral soils, on sites with similar vegetation, i.e., forested or open. So although, during the growing season, the peat soils were considerably colder than the mineral soils in the deeper layers, temperature differences between organic and mineral soils in the rooting zone were relatively small. Vegetation had a greater affect on soil temperatures in the rooting zone than did the physical make-up (mineral or organic) of the soil.

Similar results have been obtained for another peatland on the Marcel Experimental Forest. In the winter of 1968–1969, S1 Bog (Fig. 1) was strip cut, producing a pattern of alternating 30 m-wide clear-cut strips and 45 m-wide strips of 62-yr-old black spruce, oriented in an east-west direction. Crown closure of the forested strips was approximately 75%. Removal of the overstory left the clear-cut strips with a cover of Labrador tea (*Ledum groenlandicum*), leatherleaf (*Chamaedaphne calyculata*), bog laurel (*Kalmia polifolia*), blueberry (*Vaccinium angustifolium*), cotton grass (*Eriophorum tennellum*), sedges (*Carex* spp.), and sphagnum moss (*Sphagnum* spp.). Over the next 2 yr, the coverage of cotton grass and sedge increased significantly. Several times during the summer and autumn of 1969 and 1970, Verry (USDA Forest Service, North Central Forest Experiment Station, Grand Rapids, MN, unpublished data) measured soil temperatures 2–3 cm below the surface at about 50 locations distributed through the cut and uncut strips, and found the clear-cut strips to be an average of 2.6°C warmer than the forested strips in the summer and 0.7°C warmer in the autumn. The temperature difference between the forested and clear-cut strips was somewhat less in 1970 than in 1969, presumably due to the increased herbaceous cover on the cut strips in 1970.

Hayhoe et al. (1990) found that upland forests had a similar effect on soil temperatures. They compared mineral soil temperatures measured at 50 cm at a number of sites in Canada with temperatures predicted by a soil temperature model developed for soils covered by short grass (Ouellet 1973). While the model accurately predicted temperatures of cultivated soils or soils with short grass cover, forested soils were an average of 3.2°C cooler in the summer and 1.3°C cooler annually than predicted by the model.

Vegetation does not have to be of tree-height to be effective in reducing soil temperature. In the boreal forest region of western Canada, blue joint grass *Calamagrostis canadensis* often forms a dense cover following logging. Hogg and Lieffers (1991) applied experimental mowing treatments to two *Calamagrostis*-dominated sites in northern Alberta and found that a heavy accumulation of unmowed grass biomass and litter reduced average May to August soil temperatures by an average of 3.8°C at 10 cm, and delayed thawing of the soil in the spring by up to a month, compared with mowed plots.

In their soil physics text, Baver et al. (1972) state that vegetation moderates soil temperature primarily through albedo, decreased penetration of radiation through the canopy to the soil, increased latent heat loss due to evapotranspiration, and decreased heat flux to or from the soil through an insulation effect.

The cooler temperatures of forested soils cannot be accounted for by differences in albedo. Forests typically reflect less solar radiation than non-forested areas. Berglund and Mace (1972) measured albedo above the black spruce canopy in S2 Bog and the sphagnum-shrub-sedge cover of Bog Lake Fen on about 50 days from August 1968 to August 1969. The albedo of S2 Bog varied from about 6 to 8% during the snow-free season. Bog Lake Fen albedo averaged 11.6% in April and May, increased to 16.1% in June, and declined to 13.5% in October and November. The dark, uneven spruce canopy is a poor reflective surface for solar radiation compared with the sphagnum-shrub-sedge cover type. In a recent study on Bog Lake Fen, Kim and Verma (1996) found albedo to vary from 0.11 to 0.17 with a growing season midday average of 0.14 in 1991 and 0.13 in 1992. Brown (1972) reported that measurements of net radiation over a forested strip and a clear-cut strip in S1 Bog showed little difference in daytime values the first summer after logging (1969). However, the following summer, daytime net radiation over the clear-cut strip was 14% lower in June and 20% lower in September than over the black spruce strip, due to the greater albedo of the clear cut strips.

Similar results have been found for other conifer stands. McCaughey (1978) reported an albedo of 7% for a mature balsam fir and 18% after clear-cutting. Logging reduced net radiation (average of both daytime and nighttime) to the site by an estimated 10%.

Even though more radiation is absorbed by forested peatland than open peatland, the lower soil temperatures under the forest canopy indicate that less of this energy reaches the soil surface. Kim and Verma (1996) found that in Bog Lake Fen, only about 30% of the midsummer daytime net radiation was absorbed by the vascular plant canopy. The rest

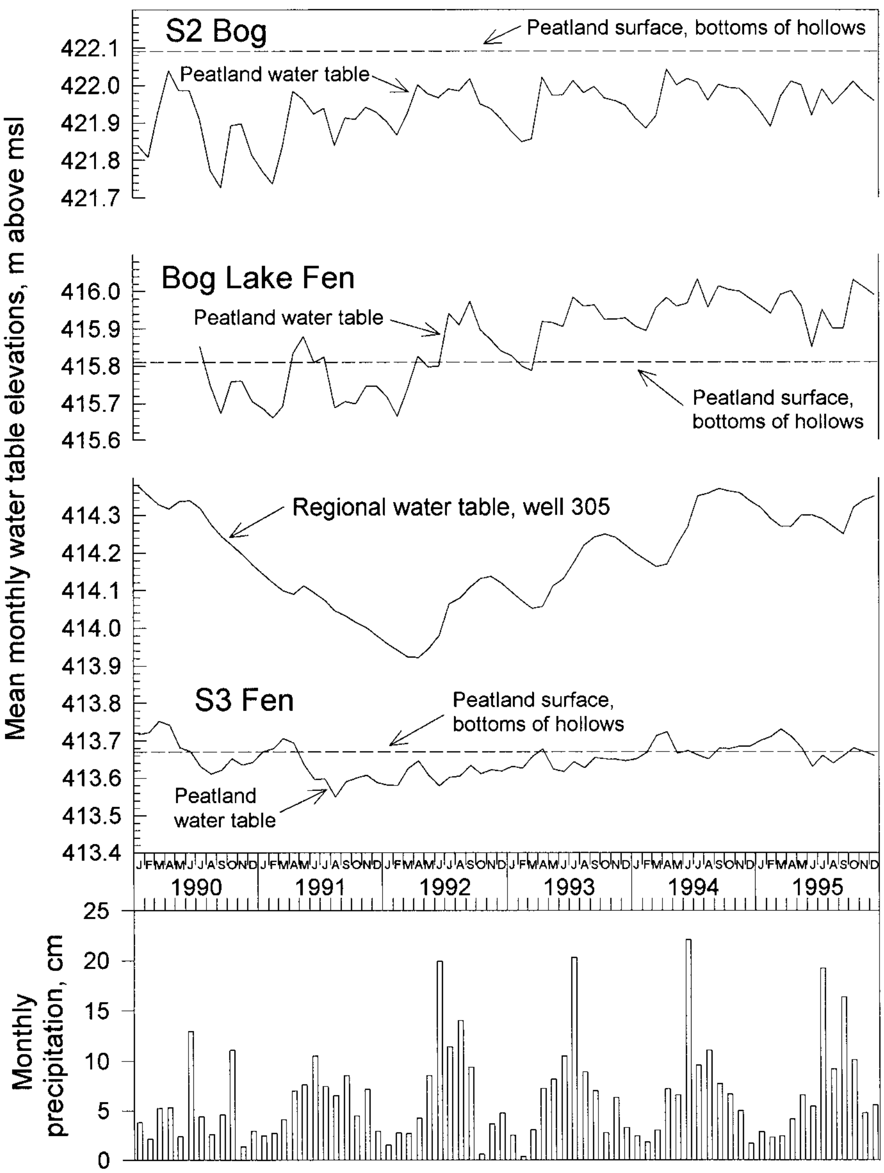


**Fig. 5**. Mean monthly elevations of water tables in S2 Bog, Bog Lake Fen, and S3 Fen, and groundwater at well 305; and monthly precipitation, Marcell Experimental Forest.

reached the sphagnum moss surface, where it was available for evaporation or conversion into sensible heat.

Balisky and Burton (1993) measured June-July temperatures in a sandy loam soil in Vancouver, British Columbia. The soil was either bare or vegetated with fireweed (*Epilobium angustifolium*), mean height 116 cm; salmonberry (*Rubus spectabilis*), mean height 56 cm; red alder (*Alnus rubra*), mean height 130 cm; or a mixture of the three species, mean height 126 cm. They found that under the various vegetative covers, mean soil temperature at 10 cm ranged from 15.2 to 19.4°C, the differences being a function of the proportion of the solar radiation that reached the ground surface.

## Precipitation and Water Level Effects

In addition to the effects of vegetation and soil type, soil temperature can also be influenced by precipitation, water table elevation, and groundwater input. Figure 5 shows mean monthly water table elevations in S2 Bog, Bog Lake Fen, and S3 Fen; mean monthly elevations of the regional groundwater table at well 305; and total monthly precipitation from 1990 through 1995.

In late autumn and winter, precipitation accumulates as snow. During this period, recharge is minimal and the regional groundwater table and water levels in the peatlands decline. In the spring, surface and shallow subsurface runoff from snowmelt plus spring rains, reaches the peatlands quickly, causing peaks in peatland water levels in March and April. Snowmelt/rain water reaches the regional water table 2 to 3 mo later, depending on the depth to groundwater, and can be seen in Fig. 5 as a rise in the water level at well 305 in May and June.

From July through October, peatland water tables and regional groundwater levels may rise or fall depending on whether growing season (May–August) precipitation is greater than or less than evapotranspiration losses. Hawkinson and Verry (1975) measured the elevation of the water table for 10 yr, at 29 wells from 2 to 20 m deep (depth to water table) in and around the Marcell Experimental Forest. They found that, on average, groundwater reaches its highest elevation for the year in early July and then declines through the rest of the summer.

Average May–August precipitation for the 1961–1995 period, measured adjacent to S2 bog, is 40.2 cm. In 1990 and 1991, May–August precipitation totaled about 22.3 and 31.1 cm, respectively. Except for the annual snowmelt peaks, regional groundwater levels declined throughout both years. From 1992 to 1994, May–August precipitation averaged 50.0 cm. In these years, groundwater levels continued to rise after the snowmelt peak each year until they began their winter decline in October or November. In 1995, May-August precipitation returned to a near-normal 40.5 cm. In this year, groundwater levels declined during the summer, before rising again in the fall in response to substantial September rains. Similar patterns can be seen in the water levels in the three peatlands.

Over the 6-yr period, water tables in S2 Bog, S3 Fen, and Bog Lake Fen averaged 15.5 cm below, 1.5 cm below, and 6.0 cm above the surface (the bottoms of the hollows) of the respective peatlands. Mean monthly water levels in S2 Bog and Bog Lake Fen varied by more than 30 cm during the 6-yr study period. The season fluctuations of the two peatlands are almost identical. The obvious difference between the two peatlands is their water levels relative to the peatland surfaces. From 1990 through 1995, monthly water levels in S2 Bog ranged from 5 to 36 cm below the surface, while monthly water levels in Bog Lake Fen varied from 15 cm below to 22 cm above the surface. The water table in Bog Lake Fen was continuously above the surface, with only hummocks extending above the water, during most of 1992 through 1995. Periodic high water levels such as these account for the lack of forest cover on Bog Lake Fen, and on the similar Junction Fen.

Due to groundwater inputs, the water table in S3 Fen is less variable, ranging from 12 cm below to 8 cm above the peatland surface during the study period. Although the water table in S3 Fen is close to the surface, this peatland is capable of sustaining a healthy black spruce forest. On average, the water level in S3 Fen is only 7–8 cm less than in the treeless Bog Lake Fen, relative to their respective surface elevations, but S3 Fen does not tend to become deeply flooded for extended periods of time in high precipitation years as Bog Lake Fen does.

The influences on soil temperatures of precipitation, peatland water tables, and groundwater vary from site to site and season to season. In spring, when the air is rapidly warming, most of the variation in soil temperature at a particular site is related to the air temperature regime. Regressions of weekly May soil temperatures at 20-cm depth at each site on mean air temperature over the 30 d and 3 d previous to each soil temperature measurement yielded $r^2$ values of 0.95 for the two upland sites and 0.74 to 0.86 for the peatlands (eqns. 1–6, Table 3). At no site did the amount of precipitation over any number of days prior to the temperature measurements have any significant effect on May soil temperatures. In the three peatlands in which water levels were recorded — S2 Bog, S3 Fen, and Bog Lake Fen — the depth of water relative to the surface of the peat did not influence soil temperatures.

From June through September, soil temperatures at all sites tended to increase with increasing precipitation amount, after the effect of air temperature was taken into account. This soil temperature response to precipitation was statistically significant at all sites, but the relative importance of precipitation amount compared with air temperature as a determinant of soil temperature varied from site to site, increasing in the order Upland-Aspen < Upland-Grass < S2 Bog < Junction Fen < S3 Fen < Bog Lake Fen. Multiple regressions of weekly 20 cm soil temperatures on mean air temperatures over the periods of 60 d and 5 to 10 d previous to each soil temperature measurement yield $r^2$ values of 0.92, 0.82, 0.82, 0.72, 0.43, and 0.55 for Upland-Aspen, Upland-Grass, S2 Bog, Junction Fen, S3 Fen, and Bog Lake Fen, respectively (eqns 7, 9, 11, 13, 15, 17, Table 3). In the regressions, the inclusion of a term for total precipitation over the 60 d previous to each soil temperature measurement increased the $r^2$ value to 0.93, 0.89, 0.91, 0.87, 0.69, and 0.85 (eqns 8, 10, 12, 14, 16, 18, Table 3). The relatively poor fit of S3 Fen soil temperature to air temperature and precipitation is due to the cooling effect of groundwater on S3 Fen in late summer, which is discussed later. For all six sites, the precipitation term that, with air temperature, resulted in the best fit with 20 cm soil temperature was total precipitation during the 60 d previous to the soil temperature measurement. Precipitation during the 15 d previous to a soil temperature measurement had no significant effect on soil temperature at any site. Although peatland water levels tend to rise and fall with variation in precipitation, soil temperatures were more closely related to actual precipitation amount rather than to water levels. When a term for water level was added to the regressions shown in Table 3, or substituted for precipitation amount, peatland water level was not a significant factor.

Figure 6 shows the relationship between precipitation and June–September 20-cm soil temperature at Bog Lake Fen. The seasonal pattern of soil temperatures calculated from 60-d and 10-d mean air temperatures (eqn. 17, Table 3) generally followed the pattern exhibited by actual measured soil temperatures (Fig. 6a). The difference between measured and calculated soil temperatures varied with precipitation. At precipitation rates significantly greater than 20 cm during the preceding 60 d, measured soil temperatures exceeded calculated temperatures, e.g. summer of 1992. During dry periods when precipitation was less than 20 cm/60 d, measured soil temperatures are significantly lower than those calculated from air temperature, e.g. summer of 1995. The pattern exhibited by the difference between actual soil temperatures at 20 cm and soil temperatures calculated from air temperature alone (Fig. 6c) is very similar to that shown by 60-d precipitation (Fig. 6b). Although water level in the peatland tended to rise and fall with long-term trends in precipitation amount, the relationship between water level and

**Table 3. Regressions of soil temperature at 20 cm depth on air temperature and precipitation amount**

| Season | Site | Regression equation | $R^2$ | |
|---|---|---|---|---|
| May | | | | |
| | Upland-Aspen | $SoilTemp_{20\text{-cm}} = 0.0 + 0.378 \times AirTemp_{30\text{-day}} + 0.333 \times AirTemp_{3\text{-day}}$ | 0.95 | (1) |
| | Upland-Open | $SoilTemp_{20\text{-cm}} = 1.2 + 0.552 \times AirTemp_{30\text{-day}} + 0.315 \times AirTemp_{3\text{-day}}$ | 0.95 | (2) |
| | S2 Bog | $SoilTemp_{20\text{-cm}} = -2.9 + 0.669 \times AirTemp_{30\text{-day}} + 0.129 \times AirTemp_{3\text{-day}}$ | 0.74 | (3) |
| | Junction Bog | $SoilTemp_{20\text{-cm}} = 0.4 + 0.594 \times AirTemp_{30\text{-day}} + 0.210 \times AirTemp_{3\text{-day}}$ | 0.79 | (4) |
| | S3 Fen | $SoilTemp_{20\text{-cm}} = -4.7 + 0.747 \times AirTemp_{30\text{-day}} + 0.278 \times AirTemp_{3\text{-day}}$ | 0.80 | (5) |
| | Bog Lake Fen | $SoilTemp_{20\text{-cm}} = -2.1 + 0.683 \times AirTemp_{30\text{-day}} + 0.332 \times AirTemp_{3\text{-day}}$ | 0.86 | (6) |
| June–Sept. | | | | |
| | Upland-Aspen | $SoilTemp_{20\text{-cm}} = 1.8 + 0.391 \times AirTemp_{30\text{-day}} + 0.337 \times AirTemp_{5\text{-day}}$ | 0.92 | (7) |
| | Upland-Aspen | $SoilTemp_{20\text{-cm}} = 1.4 + 0.389 \times AirTemp_{30\text{-day}} + 0.339 \times AirTemp_{5\text{-day}} + 0.019 \times Precip_{60\text{-day}}$ | 0.93 | (8) |
| | Upland-Open | $SoilTemp_{20\text{-cm}} = 5.4 + 0.232 \times AirTemp_{30\text{-day}} + 0.402 \times AirTemp_{5\text{-day}}$ | 0.82 | (9) |
| | Upland-Open | $SoilTemp_{20\text{-cm}} = 4.1 + 0.223 \times AirTemp_{30\text{-day}} + 0.409 \times AirTemp_{5\text{-day}} + 0.068 \times Precip_{60\text{-day}}$ | 0.89 | (10) |
| | S2 Bog | $SoilTemp_{20\text{-cm}} = 2.3 + 0.552 \times AirTemp_{30\text{-day}} + 0.094 \times AirTemp_{10\text{-day}}$ | 0.82 | (11) |
| | S2 Bog | $SoilTemp_{20\text{-cm}} = 1.0 + 0.542 \times AirTemp_{30\text{-day}} + 0.103 \times AirTemp_{10\text{-day}} + 0.074 \times Precip_{60\text{-day}}$ | 0.91 | (12) |
| | Junction Bog | $SoilTemp_{20\text{-cm}} = 5.0 + 0.289 \times AirTemp_{30\text{-day}} + 0.311 \times AirTemp_{10\text{-day}}$ | 0.72 | (13) |
| | Junction Bog | $SoilTemp_{20\text{-cm}} = 3.2 + 0.281 \times AirTemp_{30\text{-day}} + 0.333 \times AirTemp_{10\text{-day}} + 0.089 \times Precip_{60\text{-day}}$ | 0.87 | (14) |
| | S3 Fen | $SoilTemp_{20\text{-cm}} = 5.2 + 0.089 \times AirTemp_{30\text{-day}} + 0.307 \times AirTemp_{10\text{-day}}$ | 0.43 | (15) |
| | S3 Fen | $SoilTemp_{20\text{-cm}} = 2.9 + 0.048 \times AirTemp_{30\text{-day}} + 0.335 \times AirTemp_{10\text{-day}} + 0.112 \times Precip_{60\text{-day}}$ | 0.69 | (16) |
| | Bog Lake Fen | $SoilTemp_{20\text{-cm}} = 4.4 + 0.360 \times AirTemp_{30\text{-day}} + 0.282 \times AirTemp_{10\text{-day}}$ | 0.55 | (17) |
| | Bog Lake Fen | $SoilTemp_{20\text{-cm}} = 1.2 + 0.352 \times AirTemp_{30\text{-day}} + 0.312 \times AirTemp_{10\text{-day}} + 0.151 \times Precip_{60\text{-day}}$ | 0.85 | (18) |

$SoilTemp_{20\text{-cm}}$ = Soil temperature at 20 cm depth.
$AirTemp_{n\text{-day}}$ = Mean air temperature during n days preceding the soil temperature measurement (°C).
$Precip_{60\text{-day}}$ = Total precipitation during the 60 days preceding the soil temperature measurement (cm).

soil temperature was weak. In a regression of measured minus calculated soil temperature on 60-d precipitation amount, 66% of the difference between measured soil temperatures and soil temperatures calculated from air temperature was related to precipitation (Fig. 6d). A similar regression of measured minus calculated soil temperature on water table elevation relative to the peat surface yielded an $r^2$ of only 0.14 (Fig. 6e).

In the autumn, neither precipitation nor water levels had any effect on soil temperature at any site. Of the data collected in this study, October soil temperatures at 20 cm were most closely related to air temperatures during the 3 to 5 d prior to the soil temperature measurement and to the temperature of the soil at that site 4 wk before in September. Regardless of the weather, the single best predictor of October soil temperature was September soil temperature, especially in the peatlands. During the winter, any effects of peatland water levels or groundwater are minimal. Precipitation influences soil temperature mainly through the insulating ability of the snow pack.

Variations in water levels in peatlands can, at least potentially, affect soil temperatures by increasing or decreasing **evapotranspiration** (**ET**) and by changing the thermal properties — heat capacity and thermal conductivity — of the soil. In this study, however, water levels had little or no influence on peatland temperature.

During the growing season, a major portion of net radiation on peatlands is used in ET. Sensible heat flux to the soil represents only a small portion of the energy budget (Williams 1968; Stewart and Rouse 1976; Nichols and Brown 1980). Kim and Verma (1996) measured ET and energy relationships in Bog Lake Fen during May through October in 1991 and 1992. They determined that ET accounted for more than 50% of net radiation. Peatland vegetation is an efficient conveyor of water vapor to the atmosphere. Rutherford and Byers (1973) found that ET from a New Hampshire bog was 1.7 times greater than evaporation from open water. In a growth chamber study, Nichols and Brown (1980) found that water loss from 45-cm-diameter peat cores with intact peatland vegetation was twice that lost from an open water surface under the same conditions. At air temperatures from 15 to 25°C, the temperature of the water surface was about equal to air temperature, but the peat surface was 3 to 5°C less than air temperature due to evaporative cooling.

Under usual water table conditions, sufficient water is available so that peatland ET is about equal to potential ET calculated from climatic data. A substantial decrease in ET, compared with potential ET, typically does not occur during the growing season unless the water table drops to more than 30 cm below the peat surface (the peat surface being the bottoms of the hollows in a hollow/hummock topography) (Romanov 1968; Verry 1988; Kim and Verma 1996). From 1990 through 1995, the growing season water table in S2 Bog only dropped below 30 cm (varied from –30 to –40 cm) from 11 August to 2 October 1990 and from 12 August through 15 August 1991. The water tables in S3 Fen and Bog Lake Fen never fell below –30 cm during this study.

Kim and Verma (1996) partitioned ET from Bog Lake Fen into evaporation from non-vascular *Sphagnum* surfaces and transpiration from vascular plants. When the *Sphagnum* was wet, as was typical during the morning hours, about two-thirds of the water vapor flux to the atmosphere was due to evaporation. But, during sunny afternoons, even when the water table was within a few centimeters of the surface, the *Sphagnum* dried out and evaporation was sig-

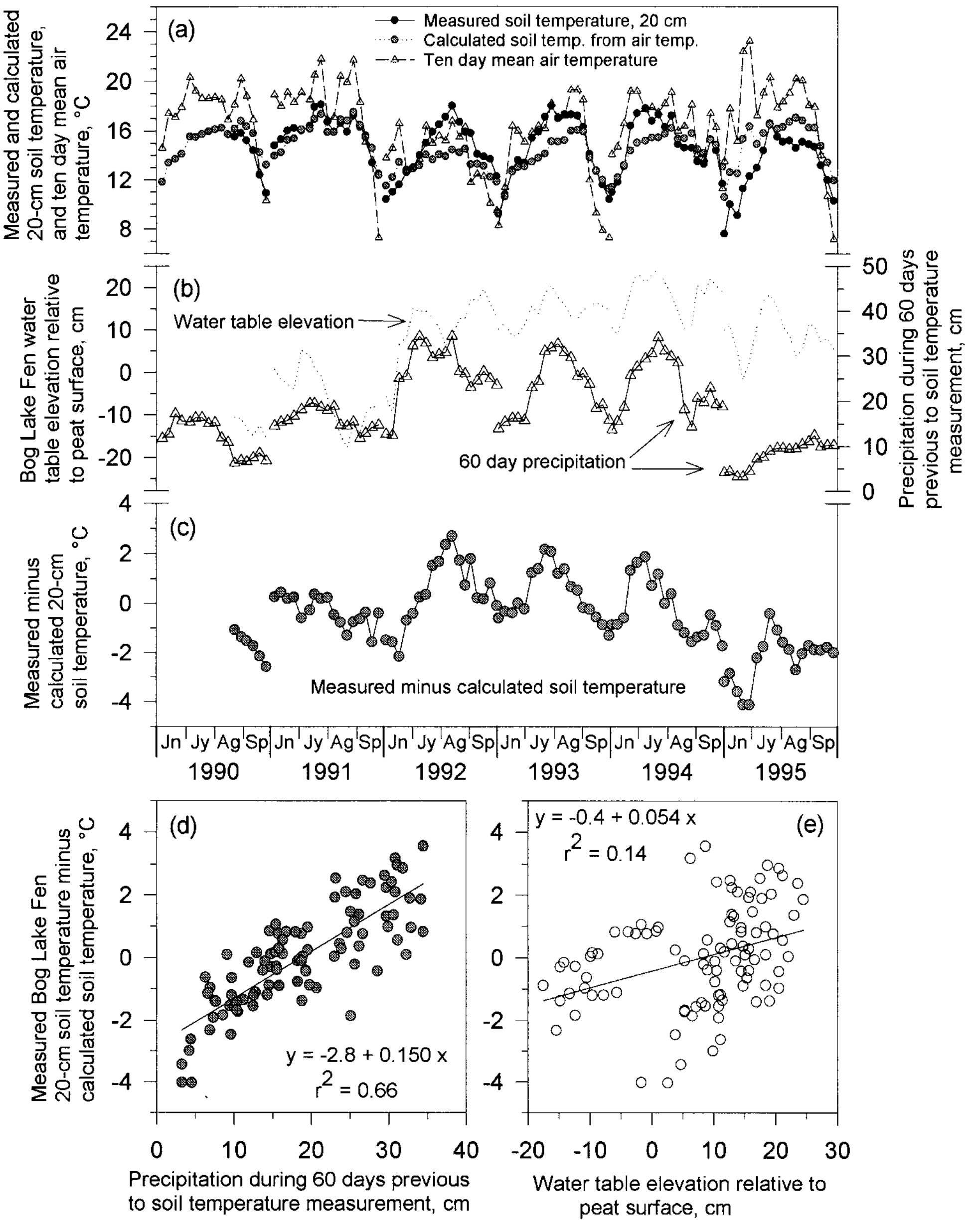


**Fig. 6**. Soil temperature, precipitation, and water table relationships in Bog Lake Fen, Marcell Experimental Forest.

nificantly diminished. However, total water vapor flux was not diminished. Sensible heat from the warming *Sphagnum* was absorbed by vascular plants, which increased transpiration and maintained ET near potential rates, as long as the water table stayed within the rooting zone (0–40 cm).

Dry peat is a good insulator. Artificial drainage of peatlands, while warming the soil surface, is often found to result in decreased rooting zone temperatures due to the greatly reduced thermal conductivity of the dry surface peat (Mannerkoski 1988; Latja and Kurimo 1988; Heikurainen and Seppälä 1964).

Midday incident solar radiation on Bog Lake Fen during the summer months (June, July, August) of 1991 averaged 670 $wm^{-2}$. That summer, precipitation was only 76% of the long-term average for the Marcell Experimental Forest. Summer precipitation in 1992 was 142% of average. Because of more frequent cloudiness, midday summertime solar radiation in 1992 averaged only 560 $Wm^{-2}$ (Kim and Verma 1996). Because of greater energy loss through ET and the development of a dry peat surface layer, warm, sunny conditions may, paradoxically, result in cooler rooting zone temperatures in peatlands, relative to air temperature, than does cloudy, rainy weather. The warming effect on peatland soil temperatures, relative to air temperature, of periods of increased precipitation seen in the present study was likely due to decreased energy loss through ET as a result of a greater frequency of cloud cover and higher relative humidity, and perhaps a greater thermal conductivity of the surface soil.

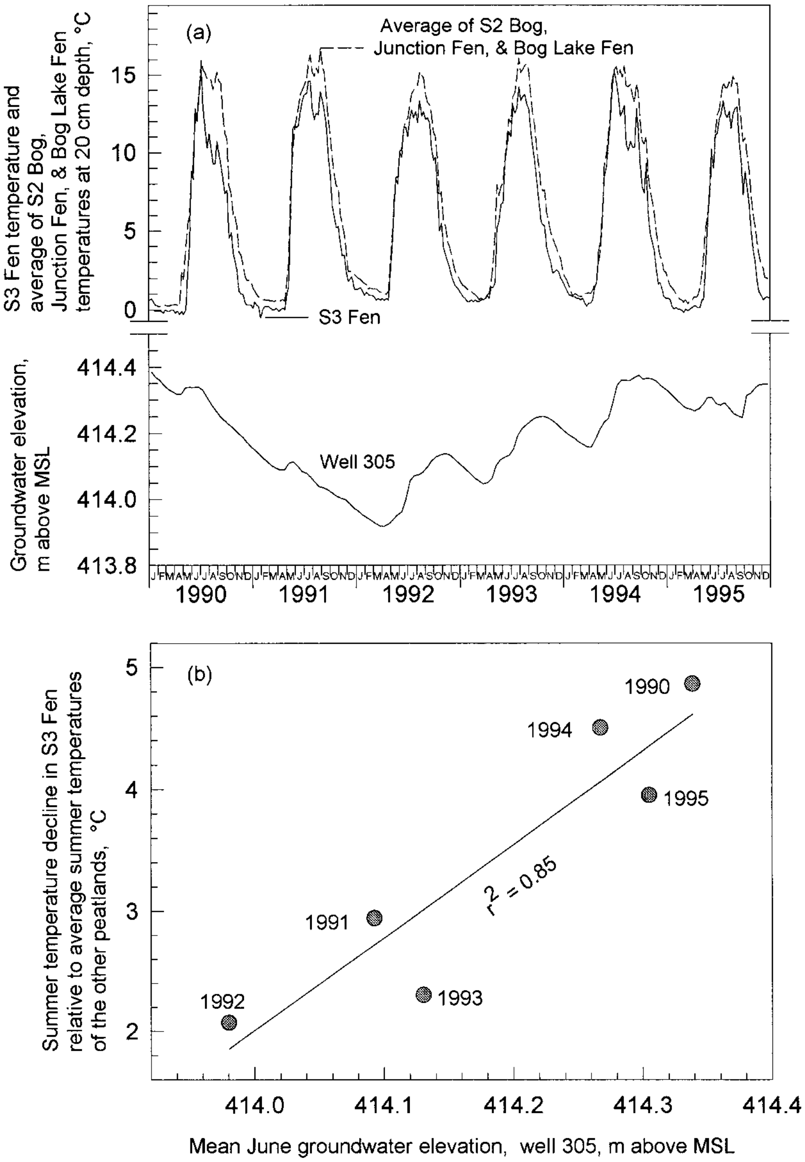


**Fig. 7**. Summer soil temperature decline in S3 Fen, relative to the average of soil temperatures in S2 Bog, Junction Fen, and Bog Lake Fen, as a function of June groundwater elevation, Marcell Experimental Forest.

## Groundwater Effects

As mentioned earlier, soil temperatures at S3 Fen were significantly cooler than at any other site because of groundwater inputs. S3 Fen intercepts the regional water table and functions as a groundwater discharge area. Evidence of groundwater input to the peatland can be seen in the chemistry of the site. The soil pH in S3 Fen, 6.0 to 6.5, compared with 4.0 to 4.5 at S2 Bog, Junction Fen, and Bog Lake Fen, is indicative of a strong influx of groundwater-borne base cations and bicarbonates. Groundwater input also influences soil temperature. Because of the cooling effects of incoming groundwater, S3 had the lowest temperature of the six sites. Mean annual soil temperature in S3 Fen was almost 1°C colder, at all depths, than temperatures at the two forested sites (S2 Bog, Upland-Aspen) and almost 2°C colder than at the open sites (Upland-Grass, Junction Fen, Bog Lake Fen).

The effect of groundwater on S3 Fen temperature can be seen in Fig. 7. Figure 7a compares weekly temperatures at 20 cm in S3 Fen with the weekly averages of 20-cm temperatures at S2 Bog, Bog Lake Fen, and Junction Fen. These are weighted averages in which S2 Bog is counted twice, because it is the only forested peatland and the other two peatlands are open. The temperature patterns at 10 and 50 cm (not shown) were similar.

Through April and May and into the early part of June each year, the temperature of S3 Fen was close to the average for the other peatlands. But, beginning generally in the latter part of June, S3 became cooler than the other sites. This condition persisted through summer and into fall and winter. The week-to-week variations in S3 temperatures in response to the vicissitudes of the weather were similar to the patterns exhibited by the other peatlands, but the actual temperature of S3 was lower than the average of the other sites. The difference between S3 and the other peatlands reached a maximum in late August to early September. Through the autumn and winter months, this difference slowly diminished, so that by the end of winter, S3 was only slightly cooler than the average of the other sites. This summer cooling of S3 is due to input of groundwater from snowmelt and spring rains. This can be seen in Fig. 7b, in which the magnitude of S3 summer cooling at 20 cm each year is regressed against the mean June groundwater elevation at well 305. The higher the groundwater table in June, the greater the input of groundwater from the snowmelt/spring rain pulse, and the greater the effect on S3 temperatures. The $r^2$ value for the regression is 0.85.

The magnitude of S3 summer cooling is calculated as the difference between S3 temperature and the weighted average temperature of the other peatland sites in late May/early June, when S3 is at its maximum with respect to the other sites, minus the difference between S3 and the other sites in late August/early September when this difference is greatest, i.e., $(S3 - \text{others})_{\text{May/June}} - (S3 - \text{others})_{\text{Aug/Sept}}$. Groundwater-induced summer cooling of S3 Fen ranged from 2.1 to 4.9°C at 20 cm.

On any summer day, at depths of 10 to 50 cm, the maximum temperature difference seen in this study among sites, i.e., the total effect of soil type, vegetation, precipitation, and groundwater, ranged from about 2 to 8°C. Other factors that were not investigated, but are likely to contribute additional soil temperature variation, include slope, aspect, soil texture, and compaction. The Marcell Experimental Forest area, like many parts of northern Minnesota, is a complex mosaic of uplands and lowlands, flat and steep, forested and open, mineral and organic soils, influenced to varying degrees by surface and subsurface water. All of these factors are potential sources of variability in soil temperature, which must be taken into account in any study or consideration of soil processes.

## CONCLUSIONS

Mean annual air temperature averaged 3.6°C. Mean annual soil temperatures at 10 to 200 cm ranged from 5.5 to 7.6°C among the six sites. Average summer soil temperatures at 20 cm ranged from 12.0 to 16.0°C among sites. Soils with open vegetation, whether mineral or peat, averaged about 1°C warmer annually and from 2 to 3°C warmer during summer than forested soils. The tall brush peatland was cooler than all other sites due to strong groundwater inputs. Mineral soils warmed more quickly in the spring, achieved higher temperatures in the summer, and cooled more quickly in the fall than peat soils; however, the greatest temperature differences between mineral and peat soils occurred at or below 50 cm. In the upper 20 cm, vegetation and groundwater had greater effects on temperature than did the physical makeup (mineral or organic) of the soil. Summer soil temperatures were higher, relative to air temperature, during periods of greater precipitation. This effect was minimal at upland sites but substantial in the peatlands. In spite of bitter cold air temperatures typical of Minnesota winters, significant frost only developed in soils in those years when severe cold weather arrived before an insulating cover of snow had accumulated.

## ACKNOWLEDGMENTS

I wish to thank Art Elling, Deacon Kyllander, Dwight Streblow, Jeff Washburn, Cindy Crawford, Mike Wakefield, Joe Huffman, Dave Stanelle, and Jack Friedline for their technical assistance in this study; and David Grigal, Shashi Verma, and Donald Baker for their critical review of the manuscript.

**Alban, D. H., Host, G. E., Elioff, J. D. and Shadis, D. 1994.** Soil and vegetation response to soil compaction and forest floor removal after aspen harvesting. USDA Forest Service Res. Pap. NC-315, North Central Forest Experiment Station, St. Paul, MN. 8p.

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

**Baver, L. D., Gardner W. H. and Gardner, W. R. 1972.** Soil physics. 4th ed. John Wiley and Sons, New York, NY. 498 pp.

**Berglund, E. R. and Mace, A. C., Jr. 1972.** Seasonal albedo variation of black spruce and sphagnum-sedge bog cover types. J. Appl. Meteorol. **11**: 806–812.

**Brown, J. M. 1972.** Effect of clearcutting on a black spruce bog net radiation. For. Sci. **18**: 273–277.

**Clement, R. J., Verma, S. B. and Verry, E. S. 1995**. Relating chamber measurements to eddy correlation measurements of methane flux. J. Geophys. Res. **100**: 21,047–21,056.

**Davis, J. R. and Lucas, R. E. 1959.** Organic soils, their formation, distribution, utilization, and management. Mich. State Univ., Agric. Exp. Stn., Spec. Bull., No. 425.

**Dise, N. B. 1992.** Winter fluxes of methane from Minnesota peatlands. Biogeochemistry **17**: 71–83.

**Ecological Stratification Working Group. 1995.** A national ecological framework for Canada. Agricultural and Agri-food Canada, Research Branch, Centre for Land and Biological Resources Research and Environment Canada, State of Environment Directorate, Ecozone Analysis Branch, Ottawa, ON. Report and national map at 1:7,500, 000.

**Hawkinson, C. F. and Verry, E. S. 1975.** Specific conductance identifies perched and ground water lakes. USDA Forest Service Res. Pap. NC-120, North Central Forest Experiment Station, St. Paul, MN. 5p.

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

**Heikurainen, L. and Seppälä, K. 1964.** The effect of drainage degree on temperature conditions of peat. Acta. For. Fenn. **76–77**: 1–30.

**Hogg, E. H. and Lieffers, V. J. 1991.** The impact of *Calamagrostis canadensis* on soil thermal regimes after logging in northern Alberta. Can. J. For. Res. **21**: 387–394.

**Kim, J. and Verma, S. B. 1996.** Surface exchange of water vapor between an open *Sphagnum* fen and the atmosphere. Boundary-Layer Meteorol. **79**: 243–264.

**Knighton, M. D. and Stiegler, J. H. 1981.** Phosphorus release following clearcutting of a black spruce fen and a black spruce bog. Pages 577–583 *in* Proceedings, Sixth International Peat Congress, 17–23 August, Duluth, MN. A. Fisher Co., Eveleth, MN.

**Latja A. and Kurimo, H. 1988.** Temperature changes in the soil and close to the ground on wetlands drained for forestry. Pages 46–51 *in* L. Rantajärvi, ed. Symposium on the hydrology of wetlands in the temperate and cold regions, 6–8 June, Joensuu, Finland, Vol. 1. Publications of the Academy of Finland 4/1988. Helsinki, Finland.

**Mannerkoski, H. M. 1988.** Effect of drainage on the temperature of surface peat. Pages 60–65 *in* L. Rantajärvi, ed. Symposium on the hydrology of wetlands in the temperate and cold regions, 6–8 June, Joensuu, Finland, Vol. 1. Publications of the Academy of Finland 4/1988. Helsinki, Finland.

**McCaughey, J. H. 1978.** Estimation of net radiation for a coniferous forest, and the effects of logging on net radiation and the reflection coefficient. Can. J. For. Res. **8**: 450–455.

**National Oceanic and Atmospheric Administration. 1993.** Climatological data, annual summary, Minnesota, 1993. National Oceanic and Atmospheric Administration, National Climatic Data Center, Asheville, NC.

**Nichols, D. S. and Brown, J. M. 1980.** Evaporation from a sphagnum moss surface. J. Hydrol. **48**: 289–302.

**Nyberg, P. R. 1987.** Soil survey of Itasca County, Minnesota. USDA Soil Conservation Service and Forest Service, Washington, DC. 197 pp.

**Ouellet, C. E. 1973.** Macroclimatic model for estimating monthly soil temperatures under short-grass cover in Canada. Can. J. Soil Sci. **53**: 263–274.

**Romanov, V. V. 1968.** Hydrophysics of bogs. Israel Program for Scientific Translations, Jerusalem, Israel. [Translated from the Russian by N. Kaner.] 299 p.

**Rutherford, R. J. and Byers, G. L. 1973.** Conserving wetland water by suppressing evaporation. Can. Agric. Eng. **15**: 9–11.

**Steel, R. G. D. and Torrie, J. H. 1960.** Principles and procedures of statistics. McGraw-Hill Book Company, New York, NY. 481 pp.

**Stewart, R. B. and Rouse, W. R. 1976.** Simple models for calculating evaporation from dry and wet tundra surfaces. Arctic Alp. Res. **8**: 263–274.

**Verry, E. S. 1975.** Streamflow chemistry and nutrient yields from upland-peatland watersheds in Minnesota. Ecology **56**: 1149–1157.

**Verry, E. S. 1988.** Hydrology of wetlands and man's influence on it. Pages 41–61 *in* Symposium on the hydrology of wetlands in temperate and cold regions, 6–8 June, 1988, Joensuu, Finland. Vol. 2. Publications of the Academy of Finland 5/1988. Helsinki, Finland.

**Verry, E. S. 1991.** Concrete frost in peatlands and mineral soils: northern Minnesota. Pages 121–141 *in* D. N. Grubich and T. J. Malterer, eds. Peat and peatlands: The resource and its utilization. Proceedings of the International Peat Symposium, 19–22 August 1991. Duluth, MN.

**Williams, G. P. 1968.** Heat balance over saturated sphagnum moss. Pages 173–193 *in* Proceedings of the 1st Canadian Congress on Micrometeorology, Ottawa, ON. February 1968, Part 1.