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Microclimatology of treeline spruce–fir forests in mountains of the northeastern United States

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Abstract

The objective of this study was to characterize the temporal and altitudinal variation in microclimate of high elevation spruce–fir (*Picea rubens–Abies balsamea*) forests in three major mountain ranges of the northeastern United States. Mean lapse rates of air temperature were comparable to those previously reported for this region, but lapse rates varied considerably in relation to diurnal (and, to a lesser degree, seasonal) effects. Mean annual soil temperatures and soil temperature heat sums did not show a consistent pattern with regard to elevation. Within our study region, it has been suggested that frequent cloud immersion at high elevation results in radiation fluxes at that are dramatically reduced compared to those at mid and low elevation, but results of this study appear not to support this hypothesis. The frequency of very high (≥90%) relative humidities increased with elevation, but although clear-sky fluxes of photosynthetically active radiation increased moderately with increasing elevation, mean mid-day fluxes during the growing season were almost identical between mid and high elevation.

Keywords: Air temperature; Lapse rate; Radiation; Soil temperature; Subalpine; Tree line

1. Introduction

Microclimatology of montane landscapes is dependent on latitude, continentality and topography. The only climatic or meteorological traits generally characteristic of montane environments are altitude-dependent decreases in atmospheric pressure (and therefore the partial pressures of O_2 and CO_2) and mean temperature (Körner, 1999). Since other meteorological parameters do not vary globally in a consistent or predictable manner, climatological assessment of a given montane location may be impossible without site-specific data (Friedland et al., 2003). Even with regional data, it has been shown that prediction of upland conditions on the basis of measured lowland conditions is easier for some variables (e.g., temperature and humidity) than others (e.g., solar radiation and potential evaporation) (Harding, 1979; Blackie and Simpson, 1993).

Mid and high elevation micrometeorological data are sparse, due primarily to difficulties of poor access and bad weather. This is certainly true in the mountains of the northeastern United States, where the cool, moist climate (Dfb by the Koppen-Geiger

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system, Reiners and Lang, 1979) is known to play an crucial role in determining the patterns of vegetational zonation, especially the transition from deciduous hardwood forest (*Betula–Acer–Fagus*) to spruce–fir coniferous forest (*Picea rubens–Abies balsamea*), and the treeline shift from spruce–fir to tundra (Botkin et al., 1972; Siccama, 1974; Reiners and Lang, 1979; Cogbill and White, 1991). Analogous montane spruce–fir forest types, with similar floristics, physiognomies and ecologies, are prevalent throughout much of the temperate zone of the Northern Hemisphere, especially eastern Asia (Reiners and Lang, 1979; Cogbill and White, 1991, discussed further in Richardson (2003)). Thus, it may be reasonable to use the high elevation spruce–fir forests of the northeastern USA as a model system for the montane spruce–fir forest type in general. However, in order to better understand the links between climate and vegetation, more climatic data from along the elevational gradient are needed. The only viable approach for fully and accurately characterizing the climate in these environments is collection of real data through field instrumentation. For example, in the northeastern USA we know little about the temporal variation (at either the diurnal or seasonal scale) in the lapse rate of air temperature, and few (if any) data have been published about elevation-related patterns in soil temperature, wind speed, or solar radiation fluxes. At a regional level, these data have additional value because the red spruce decline observed over the last four decades is thought to be related to major freezing events (Eagar and Adams, 1992; Friedland et al., 1992; DeHayes et al., 1999).

Two fully instrumented stations were established on Mt. Moosilauke, in the White Mountains of New Hampshire, and supplementary stations were established on Whiteface Mt. in the Adirondacks of New York, and Mt. Mansfield in the Green Mountains of Vermont (Fig. 1). This study improves on previous efforts in this geographic region in two regards. First, in past studies where multiple stations have been installed on a single mountain, either the research objectives were different and the elevational difference between stations was small (e.g. Friedland et al., 1992, 2003), or chart-recording thermographs (rather than electronic data loggers) were used (e.g. Siccama, 1974; Reiners et al., 1984). Second, no published studies for the northeastern USA have presented new data from multiple elevations on more than one mountain. The three mountains used here represent the three main ranges in the northeast, and offer an east–west gradient over 200 km in length.

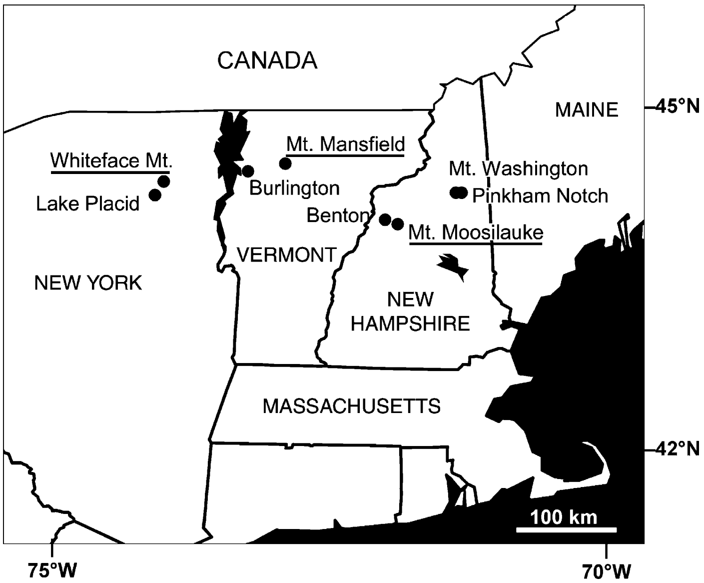

Fig. 1. Locations of meteorological stations. Underlined place names indicate stations installed by the authors specifically for this study. Additional NOAA data for Whiteface Mt. and Mt. Mansfield, as well as all other stations, were obtained from NCDC (see text for details).

2. Data and methods

Basic site data and instrumentation details are given in Tables 1 and 2, respectively, and the location of stations in relation to topography is illustrated in Fig. 2. On Mt. Moosilauke, stations were installed at mid (just below the transition from deciduous forest to spruce–fir) and high elevation (just above the treeline). Here we define treeline as the transition from a dense forest of erect stems to scattered, prostrate *krummholz* (this term, from the German for "bent wood", refers to the stunted, shrub-like form characteristic of some conifer tree species when grown under extreme environmental stress, especially at high elevation). 15-min means of air and soil temperature, relative humidity, solar radiation (in addition to quantum sensors at both elevations, a global radiation pyranometer was installed at mid elevation), and wind speed and direction were recorded. Hourly data (air temperature, relative humidity, global radiation and wind speed) from a station at the Hubbard Brook Experimental Forest, an NSF Long Term Ecological Research site managed by the United States Forest Service, were used as a low elevation station for that mountain (see http://www.hubbardbrook.org/research/data/wea/wea.htm for instrumentation details). The Hubbard Brook station is located in a valley, roughly 500 m below the lower edge of the spruce–fir zone, approximately 10 km south-east of the Mt. Moosilauke mid elevation station. On Whiteface Mt. (air and soil temperature) and Mt. Mansfield (air temperature and relative humidity), stations were installed at three different elevations within the spruce–fir zone on each mountain.

As a supplementary data source, long-term time series of monthly mean temperatures (derived from the arithmetic mean of daily maximum and minimum temperatures) were obtained from NOAA's National Climatic Data Center (NCDC), and the State University of New York's Atmospheric Sciences Research Center (Wilmington, NY), for seven additional stations (Fig. 1). Lapse rates were calculated with these data using paired stations: Mt. Washington (1900 m asl)–Pinkham Notch (612 m asl), Mt. Mansfield (1204 m asl)–Burlington Airport (101 m asl), and Whiteface Mt. (1485 m asl)–Lake Placid (591 m asl).

Table 1
Meteorological station locations and abbreviated site descriptions

Low	Mid	High
Adirondack Mts.: Whiteface Mountain (summit 1485 m asl, 44°22′N, 73°54′W)		
1095 m	1377 m (+ 282 m)	1475 m (+ 380 m)
SE aspect	S aspect	S aspect
Edge of ski trail	Transition from closed forest	Scattered *krummholz*, talus
Open to S, closed spruce-fir forest to N	Trees ≈2–3 m in height	Flagged stems ≈1 m in height
Trees 12–16 cm DBH	Trees 12–14 cm DBH	
Green Mts.: Mount Mansfield (summit 1339 m asl, 44°33′N, 72°49′W)		
917 m	1197 m (+280 m)	1317 m (+400 m)
SE aspect	Flat, ridgetop	Flat, ridgetop
Edge of toll road to summit	Just below transition to *krummholz*	Scattered *krummholz*
Closed spruce-fir forest to N and S	Trees ≈4 m in height	Flagged stems ≈1 m in height
Trees 18–20 cm DBH	Trees 7–11 cm DBH	
White Mts.: Mount Moosilauke (summit 1463 m, 44°01′N, 71°51′W)		
247 m	748 m (+501 m)	1425 m (+1178 m)
Flat site on valley floor	Flat to moderately SW aspect	Flat to moderately E aspect
Mowed clearing, hardwood forest	In broad Baker River valley	Immediately above treeline, exposed
Hubbard Brook research station	Forest and ridge directly to east	Near top of Moosilauke East Peak
	Just below lower edge of spruce-fir	Larger stems in sheltered hollows

Elevations given in parentheses (e.g. Whiteface mid, +282 m) denote elevation difference between "mid" or "high" elevation stations and "low" elevation stations.

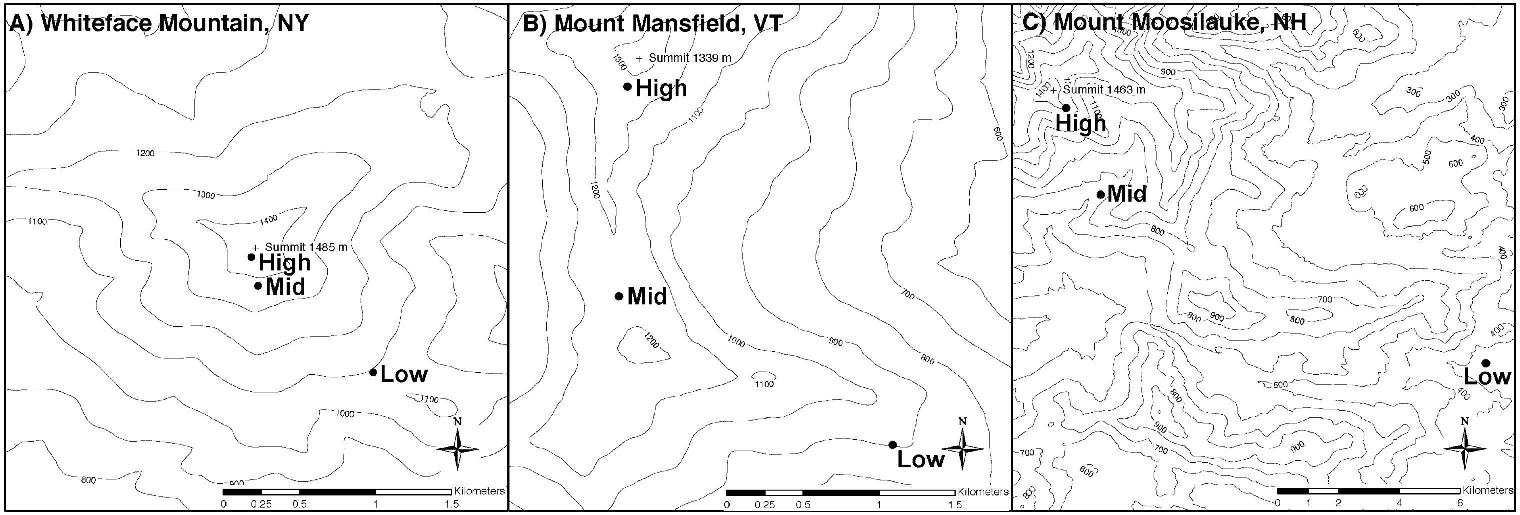

Fig. 2. Meteorological station locations in relation to mountain topography. Elevations are in m asl. Scale is the same in (A) Whiteface Mt. and (B) Mt. Mansfield, but reduced in (C) Mt. Moosilauke in order to show location of low elevation station in the Hubbard Brook valley.

Table 2
Instrumentation details for meteorological stations installed for this study

Whiteface Mt. (June 2001–September 2002): 1095, 1377, 1475 m
Hobo H8 Pro (Onset Computer Corp., Bourne, MA) data logger with integral thermistor for air temperature, and external probe for soil temperature (15 cm depth), housed in a vented radiation shield
30 min sampling interval
Mt. Mansfield (October 2000–September 2002): 917, 1197, 1317 m
Hobo H8 Pro data logger with integral thermistor for air temperature and capacitive sensor for relative humidity, housed in a vented radiation shield
30 min sampling interval
Mt. Moosilauke (September 2001–September 2002): 748, 1425 m (data from a station at 247 m in the Hubbard Brook Experimental Forest were used as a low elevation station, see http://www.hubbardbrook.org/research/data/wea/wea.htm for instrumentation details)
Multi-channel data logger (CR-10, Campbell Scientific Inc. [CSI], Logan, UT)
Copper–constantan thermocouples (in conjunction with CR10XTCR (CSI) reference thermocouple) for air temperature and soil temperature (15 cm depth)
Capacitive sensor for relative humidity (CS 500 at 748 m, HMP35C at 1425 m; both CSI), mounted (along with air temperature thermocouple) in a vented radiation shield
Quantum sensor (190SZ, Li-Cor Inc., Lincoln, NE at both 748 and 1425 m) and pyranometer (Model 50, Eppley Laboratory, Newport, RI at 748 m elevation only)
Wind speed and direction (03001 Wind Sentry, R.M. Young Co., Traverse City, MI)
10 s sampling interval, 15-min means output to storage module

3. Results and discussion

3.1. Air temperature

The location of the alpine treeline is generally considered to be driven by growing season air temperatures, although small-scale variation in treeline elevation may be due to a multitude of other factors, including topography, aspect, wind and winter snow accumulation (Griggs, 1938; Daubenmire, 1954; Tranquillini, 1979; Cogbill and White, 1991; Körner, 1999). In the present study, mean July air temperature at treeline was 13 °C on both Whiteface Mt. and Mt. Moosilauke, and 14 °C on Mt. Mansfield. Within the Appalachian Mountain chain, Cogbill and White (1991) reported that treeline correlated with the 13 °C July isotherm, whereas Daubenmire (1954) found that treeline locations in western North America correlated well with the 10 °C July isotherm. Even taking in to account the fact that, in the northeastern United States, July 2002 was roughly one-half a degree warmer than normal (based on the 1961–1990 station norms for the six NOAA stations used here), these data support the idea that air temperatures at treeline in this region are warmer than those in the western USA. Treeline thus occurs at a lower elevation than would be expected on the basis of air temperature alone. This is especially true for Mt. Mansfield, where the treeline ecotone is located at an elevation almost 200 m lower than on either of the other two mountains. Other factors, such as wind and snow blast (Cogbill and White, 1991) could be the cause of this treeline depression. However, it must also be remembered that treeline is a "mobile migration front" rather than a "static climatic boundary" (Griggs, 1934). Because trees are typically long-lived, slow to establish, and somewhat resilient to environmental change, changes in treeline position usually lag behind changes in climate (Körner, 1999). In other words, treeline position represents the long-term average (i.e. decades to centuries) climatic history of a site. When there is an upward time trend in temperatures (i.e., global warming—mean annual air temperatures in New Hampshire, for example, increased by about 1 °C between 1895 and 1999, see New England Regional Assessment Group, 2001), treeline should be found at an elevation lower (i.e., warmer) than expected, because the upslope movement of trees cannot take place instantaneously.

Mean annual temperatures (October 2001–September 2002 data using stations installed for this study as

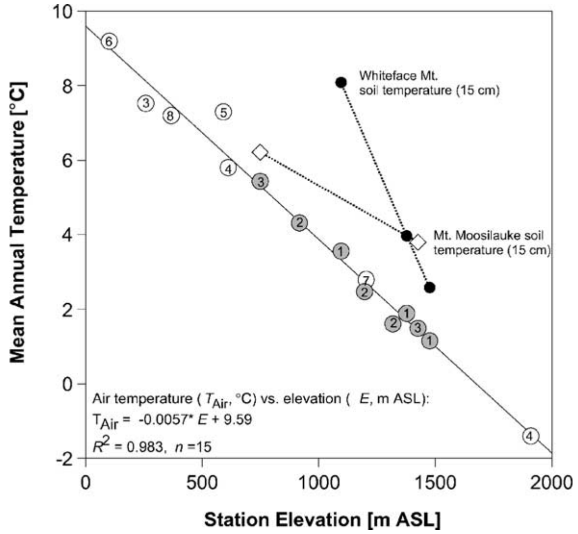

Fig. 3. Mean annual temperature (based on October 2001–September 2002 measurements) plotted against station elevation for selected weather stations in the northeastern United States. Numbers inside circles refer to sites: (1) Whiteface Mt., NY; (2) Mt. Mansfield, VT; (3) Hubbard Brook and Mt. Moosilauke, NH; (4) Mt. Washington, NH and Pinkham Notch, NH; (5) Lake Placid, NY; (6) Burlington Airport, VT; (7) Mt. Mansfield, VT; and (8) Benton, NH. Shaded symbols for sites 1–3 indicated data collected by the authors. NOAA data for sites 4–8 were obtained from the National Climatic Data Center (see text for further details). Soil temperature gradients on Whiteface Mt. and Mt. Moosilauke are shown for comparative purposes. The linear regression line is based only on air temperature data.

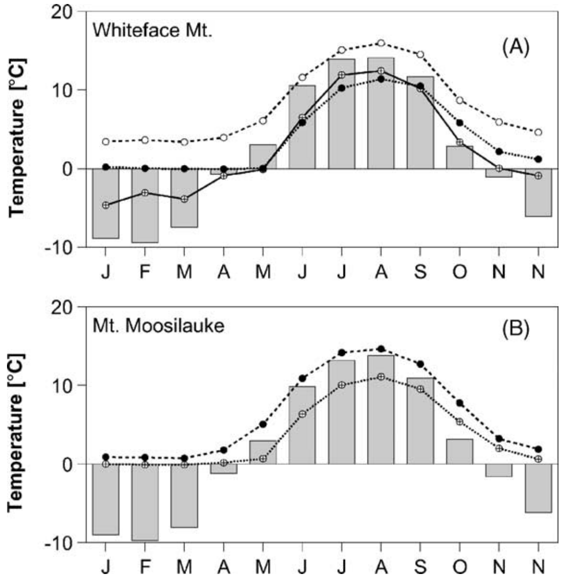

Fig. 4. Monthly mean air and soil temperatures (October 2001–September 2002) on: (A) Whiteface Mt. and (B) Mt. Moosilauke. Shaded bars represent air temperature at treeline stations (1377 and 1425 m asl, respectively); soil temperatures are shown as line plots [(○) low elevation; (●) mid elevation; (⊕) high elevation].

well as NOAA stations) exhibited a very direct relationship with elevation ($R^2 = 0.99$, $n = 15$, Fig. 3), despite the fact that stations are spread out across a wide geographic range (200 km from east to west) and presumably have different climatic influences. Multiple regression analysis indicated that mean annual temperature decreased at a rate of $0.57 \pm 0.02\,^{\circ}$C per 100 m elevation (coefficient significantly different from 0 at $P \leq 0.001$), whereas it increased at a rate of $0.29 \pm 11\,^{\circ}$C per degree of longitude from east to west ($P = 0.03$). Temperature was not correlated with latitude, presumably because the narrow latitudinal band across which the stations were located (less than 1°) made such patterns difficult to detect. In contrast, Leffler (1981) calculated that mean annual temperature decreases by 1.08 °C per degree of latitude along the Appalachian Mts. Although Leffler (1981) and Schmidlin (1982) demonstrated similar elevation–temperature relationships for mountaintop, or "crest" stations, the present data include a number of stations located either at mid-slope or in valleys, and thus validate a much more general relationship across a greater variety of topographical situations.

Monthly mean air temperatures at different elevations on the same mountain were very well-correlated with each other, and seasonal patterns were similar across the three different mountains (Fig. 4). Mean annual lapse rates on each mountain were intermediate between the dry adiabatic lapse rate (0.98 °C/100 m) and the saturated adiabatic lapse rate (0.50 °C/100 m at 20 °C) (Barry, 1992), and compared favorably with those previously reported for the eastern United States (Table 3), as well as mountain ranges in Japan and Europe (Barry, 1992; Rolland, 2003) and the tropics (Körner, 1999). These figures were also in keeping with (though slightly steeper than) those calculated using NOAA monthly mean temperature data (Table 3). There was a longitudinal trend to the lapse rates at the stations installed for this study, with the most easterly mountain (Mt. Moosilauke) having the least steep lapse rate, and the most westerly mountain (White-

Table 3
Comparison of air temperature lapse rates (°C/100 m) calculated in this study with those previously published for the mountains of the eastern United States

Lapse rate	Location	Source
0.34	Mt. Ascutney, VT (winter only)	Friedland et al. (2003)
0.41	Great Smoky Mountains, NC-TN	Shanks (1954)
0.5	White Mountains, NH	Sabo (1980)
0.53	Mt. Mansfield, VT	NOAA data (1957–2001)
0.56	Mt. Washington, NH	NOAA data (1950–2001)
0.57	Whiteface Mt., NY	NOAA data (1985–1988)
0.57	NY, VT and NH data	Fig. 3, this study
0.58	Mt. Moosilauke, NH	This study
0.58	Mt. Washington, NH	Leffler (1981)
0.60	22 NOAA sites, VT and NH	Dingman (1981)
0.6	Camels Hump, VT	Siccama (1974)
0.62	Mt. Mansfield, VT	This study
0.64	Mt. Moosilauke, NH (April–November only)	Reiners et al. (1984)
0.64	Whiteface Mt., NY	This study
0.70	Whiteface Mt., NY	Miller et al. (1993b)

face Mt.) having the steepest lapse rate. This pattern is at least consistent with the idea that the Adirondacks have a drier, more continental climate than the White Mountains, which are more humid and maritime (Miller et al., 1993b).

Both Fig. 3 and Table 3 are based on annual means, and therefore obscure the considerable variation in lapse rates that occurred at shorter temporal scales. For example, the standard deviation of the mean annual lapse rate calculated using NOAA data was ±0.04 °C/100 m ($n = 101$ station-years), and the standard deviation of the mean monthly lapse rate was ±0.10 °C/100 m ($n = 1175$ station-months). There was a general tendency for lapse rates to be least steep during the autumn (October–December) and winter (January–March), and most steep during the spring (April–June) and summer (July–September) (Table 4). Although these results fit in with what is generally observed in temperate zone mountains (Körner, 1999; Rolland, 2003), the pattern was by no means universal. For example, Whiteface Mt. data from this study exhibited little or no seasonality (Whiteface Mt. data from NOAA exhibited only modest seasonality), and Siccama (1974) reported that lapse rates on Camels Hump were steepest in February and March (0.8 °C/100 m) and least steep in July and August (0.4 °C/100 m).

At shorter time scales, the lapse rate variability was even more pronounced (Table 4). For example, the mean lapse rate on Mt. Moosilauke, calculated using 15-min means, was 0.58 °C/100 m, but the standard deviation was 0.41 °C/100 m. Thus, 50% of all values lay within the range 0.37–0.87 °C/100 m. However, the lapse rate was very strong (≥1.0 °C/100 m) for more than 10% of the time. The distribution of lapse rates was skewed (non-normal), and this skew was more pronounced during the months October–March than April–September (Fig. 5). Inversions (i.e. negative lapse rate) were observed only slightly less often during the autumn and winter (9.4% of all observations) than during the spring and summer (10.8% of all observations).

There was a clear diurnal lapse rate pattern, with lapse rates during the day generally being much stronger than those at night (Fig. 6). The diurnal variation in lapse rate was stronger on Mt. Moosilauke than the other two peaks, and this was driven by elevation

Table 4
Quarterly mean lapse rates of air temperature (°C/100 m elevation) on mountains in the northeastern United States

	JFM	AMJ	JAS	OND
Calculated using half-hourly or quarter-hourly data (Authors' stations)				
Whiteface Mt., 1095–1475 m asl (10/01–10/02)	0.64 ± 0.61	0.63 ± 0.32	0.65 ± 0.30	0.63 ± 0.37
Mt. Mansfield, 917–1317 m asl (10/00–10/02)	0.56 ± 0.54	0.65 ± 0.36	0.64 ± 0.27	0.64 ± 0.39
Mt. Moosilauke, 748–1425 m asl (9/01–10/02)	0.60 ± 0.35	0.63 ± 0.43	0.58 ± 0.43	0.52 ± 0.42
Calculated using monthly mean data (NOAA/ASRC stations)				
Mt. Washington, 612–1900 m asl (1950–2001)	0.49 ± 0.06	0.63 ± 0.03	0.59 ± 0.03	0.51 ± 0.05
Mt. Mansfield, 101–1204 m asl (1957–2001)	0.45 ± 0.10	0.57 ± 0.07	0.56 ± 0.05	0.52 ± 0.06
Whiteface Mt., 591–1485 m asl (1985–1988)	0.54 ± 0.06	0.56 ± 0.02	0.60 ± 0.10	0.58 ± 0.06

JFM, January–March; AMJ, April–June; JAS, July–September; OND, October–December. Values are mean ± 1 S.D.

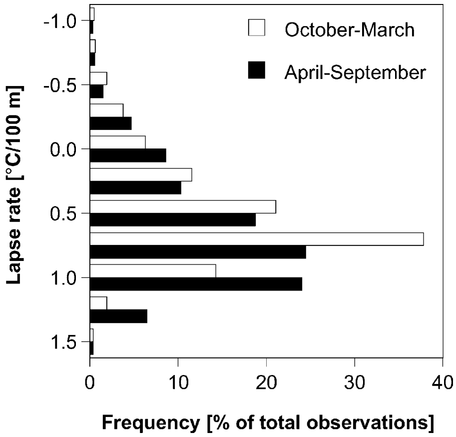

Fig. 5. Histogram of air temperature lapse rates calculated between mid (748 m asl) and high elevation (1425 m asl) stations on Mt. Moosilauke (September 2001–September 2002).

differences in the diurnal pattern of air temperature, as well as the magnitude of the diurnal temperature range (DTR = mean[daily maximum − daily minimum]), on that mountain. On Mt. Moosilauke, DTR was larger at low (10.7 ± 4.8 °C) and mid (9.4 ± 4.1 °C) elevation than at high elevation (6.9 ± 3.1 °C). Although this fits the expected pattern (Körner, 1999), on both Whiteface Mt. and Mt. Mansfield, there was no elevation pattern in the mean diurnal temperature range, which was fairly constant across elevations at about 7.5 °C. The phase, or timing, of the diurnal lapse rate pattern also differed among mountains. For example, on Mt. Moosilauke, lapse rates reached a minimum at 0600 h and peaked at 1500 h. In contrast, on Whiteface Mt. and Mt. Mansfield, lapse rates were quite stable from 1800 to 0600 h, but reached a maximum in mid-morning. Both magnitude and phase differences in the diurnal pattern can likely be explained by the different topographies of the three mountains (Fig. 2). For all three of the mountains, high elevation sites were located in exposed positions where strong winds and good mixing were probably common. The low elevation sites on Whiteface Mt. and Mt. Mansfield were located mid-slope, which contrasts with the mid elevation site on Mt. Moosilauke, which was located in a valley. Valley sites, which typically have strong diurnal patterns, warm more during the day, and cool more during the night, than mid-slope sites (Barry, 1992). The diurnal mountain/valley breeze pattern (described below), as well as differences in aspect, may further enhance this effect and contribute to the diurnal lapse rate pattern. For example, night-time cold air drainage into a valley would reduce the valley-to-summit temperature differential, and hence the lapse rate.

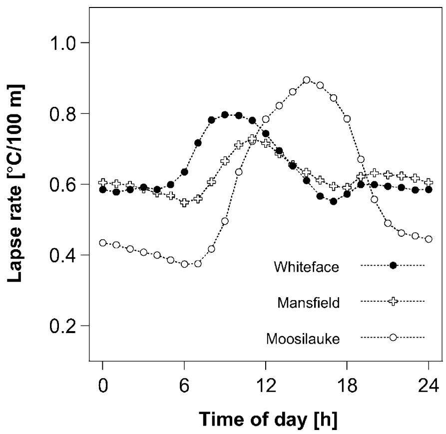

Fig. 6. Diurnal variation in the lapse rate of air temperature on Whiteface Mt., Mt. Mansfield, and Mt. Moosilauke. On Whiteface Mt., lapse rate calculated between high (1475 m asl) and low (1095 m asl) elevation stations. On Mt. Mansfield, lapse rate calculated between high (1317 m asl) and low (917 m asl) elevation stations. On Mt. Moosilauke, lapse rate calculated between high (1425 m asl) and mid (748 m asl) elevation stations.

3.2. Soil temperatures

In terms of mean annual soil temperature, the low elevation site on Whiteface Mt. appears to be an outlier in Fig. 3. And, for monthly means, the elevation trend on Whiteface Mt. was not entirely consistent. For example, at mid elevation, soil temperatures were warmer than those at high elevation, except during the months June, July and August, when they were cooler (Fig. 4). Thus, compared to air temperatures, soil temperatures did not vary as consistently with elevation. In comparison, in the Great Smoky Mountains of the Southern Appalachians, Shanks (1956) found that the elevational gradient for mean soil temperature (at 15 cm depth) was similar to that for air temperature from May to October. Similarly, Siccama (1974) reported that mean annual soil temperature (15 cm) on Camels Hump, Vermont, decreased linearly with elevation, from 7.2 °C at 549 m to 3.9 °C at

1158 m (0.54 °C/100 m). In a European survey, Green and Harding (1980) found that the altitudinal gradient of soil temperature varied predictably throughout the year, with the gradient largest in summer (to about 1 °C/100 m elevation), and smallest (close to 0) in winter. Although the Mt. Moosilauke data fit this pattern, the Whiteface Mt. data do not.

In the present study, soils were frozen for much of the year at the higher elevation sites (10 weeks of the year at mid elevation, and 28 weeks of the year at high elevation, on Whiteface Mt., and 17 weeks of the year at high elevation on Mt. Moosilauke; this in spite of the fact that between October 2001 and March 2002, mean air temperatures across the region were about 2.7 °C warmer than normal). In comparison, Siccama's (1974) study reported that, at a depth of 15 cm, soils on Camels Hump, Vermont, froze for only a few weeks of the year, even at the highest elevation sites (1158 m).

There are a number of possible explanations for our apparently anomalous results. The inversion of the elevation–temperature relationship during summer months on Whiteface Mt. may be related to the open spacing and sparse crowns of trees at high elevation, which allows more solar radiation to reach the soil surface compared to mid elevation (Körner, 1999). A contributing factor may be that windswept *krummholz* sites can have surprisingly thin snow cover compared to lower elevation sites, although weekly measurements of snowpack depth would be required to verify this conjecture. A thin layer of snow provides less insulation against extreme cold in the winter, and melts more quickly in the spring, compared to a thick layer. At high elevation, monthly mean soil temperature dropped well below freezing during the winter months, whereas at mid elevation, monthly mean soil temperature remained right at freezing, and at low elevation, monthly mean soil temperature never dropped below 3 °C. On Mt. Moosilauke, high elevation soil temperatures remained right at freezing throughout the winter months, whereas this threshold was never reached at mid elevation.

On Mt. Moosilauke, additional soil temperature probes were installed at sites approximately 50–100 m from each of the main weather stations, thus providing a measure of the variation in soil temperature at a given elevation. At both high and mid elevation sites, mean annual soil temperature measured by the supplementary probe was within 0.2 °C of that measured by the primary station. These data suggest that soil temperatures measured on Mt. Moosilauke are representative of their sampling elevation.

3.3. Air and soil temperature thresholds: heat sums

Many physiological processes show a temperature response (e.g. photosynthesis and respiration, see Tranquillini, 1979), and various temperature thresholds have been suggested for different aspects of plant function. For example, soils must be unfrozen if water uptake by roots is to occur, whereas air temperatures in the range 5–7 °C are thought necessary for the development of new tissues (this is the basis of Körner's (1999) "sink oriented hypothesis" for treeline location, which was recently supported by the work of Hoch et al. (2002)). In this study, mean air temperatures were above freezing for about 25–30 weeks of the year at all elevations and on all mountains. There was a near-perfect linear relationship between station elevation and heat sums at all three reference temperatures used (0, 5 and 10 °C). For example, with a reference temperature of 0 °C, the annual heat sum ($H_{0\,^\circ\mathrm{C}}$, with units of degree days), could be predicted for a station at elevation E (m asl) by:

$$H_{0\,^\circ\mathrm{C}} = 3595 - 1.25E \quad (R^2 = 0.99, n = 9)$$

This relationship compares quite favorably with that derived by Reiners et al. (1984), who reported that $H_{0\,^\circ\mathrm{C}}$ decreased at a rate of 1.3 degree-days/m elevation. The present heat sum–elevation relationship became less steep as the reference temperature increased. For example, $H_{5\,^\circ\mathrm{C}}$ decreased at a rate of 0.95 degree-days/m (intercept = 2343, R^2 = 0.99), whereas $H_{10\,^\circ\mathrm{C}}$ decreased at a rate of 0.71 degree-days/m (intercept = 1370, R^2 = 0.99). Forest stand modeling at Hubbard Brook by Botkin et al. (1972) has suggested that heat sums may be the critical factor in determining the vegetational zonation along elevational gradients in the northeastern United States.

Soil temperature heat sums did not follow the same linear pattern. For example, on Whiteface Mt., the heat sums calculated with reference temperatures of 5 and 10 °C were higher at the high elevation site (636 and

127 degree-days) than the mid elevation site (571 and 54 degree-days), which is the opposite of the expected pattern. Coupled with the somewhat anomalous soil temperature data from the low elevation station on Whiteface Mt. (Fig. 3), these data from the mid and high elevation sites further indicate that strict correlations between elevation and soil temperature do not always occur.

3.4. Relative humidity

Cooler temperatures and frequent cloud immersion (Siccama, 1974) result in high relative humidities on mountain summits in this region. On Mt. Moosilauke, the frequency of very high relative humidities (RH $\geq$ 90%) increased with station elevation (Fig. 7), and similar results were obtained for Mt. Mansfield (data not shown). Across the whole year, RH $\geq$ 90% was more than twice as common at high elevation as at low elevation (Fig. 7). However, the difference between low and high elevation was less extreme during the summer months (July–September) compared to the winter months (January–March). This may be partially due to periodic rime icing of the high elevation radiation shield, which could inflate the frequency of very high humidities recorded in winter. In the winter, RH was 90% or greater for 27% of observations at the low station, 65% of observations at the mid station, and 75% of the observations at the high station.

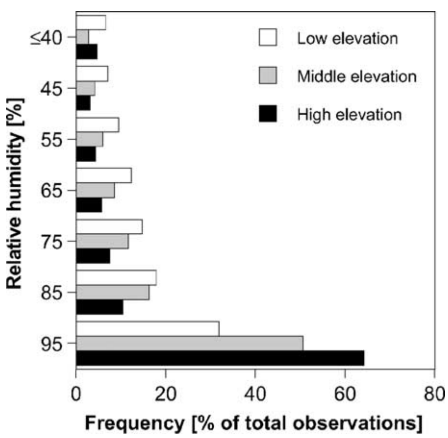

Fig. 7. Histogram depicting frequency distribution of relative humidity observations at three elevations on Mt. Moosilauke (October 2001–September 2002).

In the summer, RH was 90% or greater for 36% of observations at the low station, 40% of the observations at the mid station, and 50% of the observations at the high station.

3.5. Wind speed and direction

At high elevation on Mt. Moosilauke, measured wind gusts reached a maximum of 22 m/s during the period June–September 2002 (the only period for which consistently valid data were obtained, due to rime icing of the anemometer cup in cold weather). In contrast, at mid elevation, gusts reached only 12 m/s. Actual gusts may have been higher, as both the inertia of the cup anemometer and the 10 s sampling frequency could result in some damping of the measured peak. Mean wind speeds decreased with decreasing elevation, from 4.8 m/s at high elevation to less than 1.0 m/s at mid and low elevations (Fig. 8). A reasonable estimate of winter wind speeds can be obtained by doubling the above figures: the mean wind speed on the summit of Mt. Washington is 12 m/s during the summer and nearly twice that (23 m/s) during the winter (Barry, 1992).

The data were reasonably well-modeled using a Weibull distribution (Lee, 2000), and both the scale factor *c* and shape factor *k* differed across elevations (Fig. 8). Thus, extreme wind events were much less common at both low and mid elevation than at high elevation. For example, at high elevation, 19% of 15-min mean wind speeds were greater than 6 m/s, whereas at mid and low elevations, the wind was never as consistently strong. Topography (i.e. the exposed position of the high elevation station on the East Peak of Mt. Moosilauke), rather than elevation, is thought to be the key factor that determines mountain wind patterns (Barry, 1992). This hypothesis is supported by the patterns in wind direction observed on Mt. Moosilauke. At the high elevation station, there were no diurnal patterns in wind direction (Fig. 9), as the prevailing winds were from the SW and NW, regardless of time of day. In contrast, at the mid elevation station there was a classic, clearly defined valley wind (upslope or anabatic) during the day and mountain wind (downslope or katabatic) at night. At the mid elevation site, the Baker River valley runs in a south-west to north-east direction, which almost perfectly matches the observed directional pattern. During the day, winds

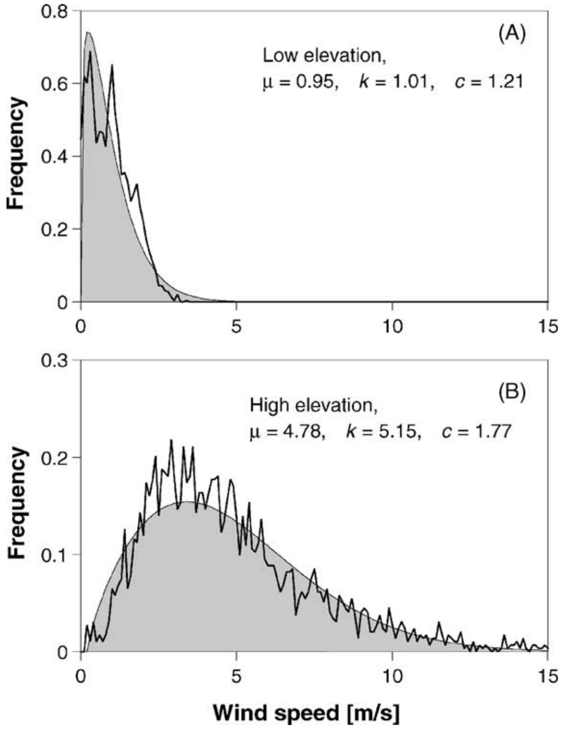

Fig. 8. Plot of frequency distribution of 15-min mean wind speed observations at two elevations on Mt. Moosilauke (June 2002–September 2002). The modeled Weibull distribution (with shape, c, and scale, k, parameters as indicated) is illustrated for comparative purposes. The overall mean wind speed is denoted by μ. (A) Low elevation, 247 m asl and (B) high elevation, 1425 m asl.

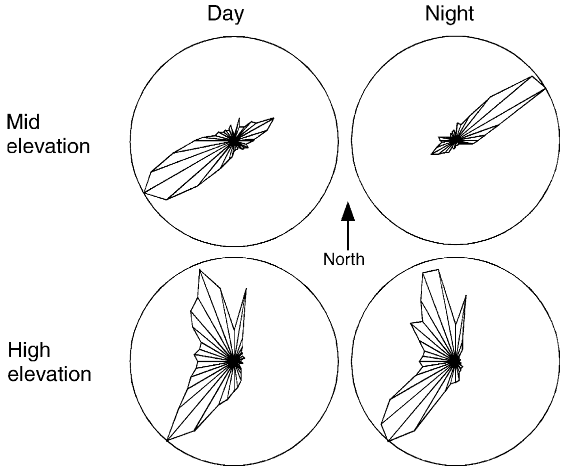

Fig. 9. Polar histograms depicting the frequency distribution of wind direction in relation to time of day at two elevations on Mt. Moosilauke (June–September 2002). The data indicate a clear mountain/valley breeze system at mid elevation (748 m asl), but no diurnal pattern at high (1425 m asl) elevation.

blew from southwest, up the valley to the northeast. The pattern was reversed at night.

Miller et al. (1993b) reported that mean above-canopy wind speeds during the growing season on Whiteface Mt. ranged from ≤1 m/s at 600 m to 5 m/s or greater at 1350 m. These values are in keeping with those reported here. The Moosilauke data fit the semi-exponential elevational pattern shown by Miller et al. (1993b) almost perfectly. In contrast, figures by Sabo (1980) suggest that mean above-canopy wind speeds in the White Mountains increase from 2.7 m/s at 500 m, to 8.3 m/s at 1000 m and 12.5 m/s at 1500 m. The Sabo values are much higher than ours, but because the data source or station locations are not reported, it is difficult to hypothesize about the factors responsible.

3.6. Global radiation and PPFD

On clear days at Mt. Moosilauke, mid-day global radiation fluxes (R) at mid elevation were about 5% higher (10% per 1000 m) than those at low elevation, and PPFD fluxes (Q) at high elevation were about 4% higher (7% per 1000 m) than those at mid elevation. This agrees with the general result that solar radiation increases with increasing elevation (with the effect most pronounced below 2000 m) as a result of a shorter atmospheric path length, which reduces molecular scattering and absorption by gases (Barry, 1992). However, our values are lower than the rate (15% per 1000 m) reported by Harding (1979) for Scotland's Cairngorm. Day-to-day variation in this rate is in large part determined by the amount and size distribution of aerosols in the lower atmosphere (Harding, 1979), and so overall between-site differences could be attributed to the same factors.

When cloudy days are included and data are averaged on a monthly basis, results were somewhat different. Monthly mean mid-day R was somewhat higher at low elevation on Mt. Moosilauke compared to mid elevation (Fig. 10A). This pattern was more or less consistent across the calendar year, and can probably be attributed to the greater frequency of local cloud immersion (which blocks incoming radia-

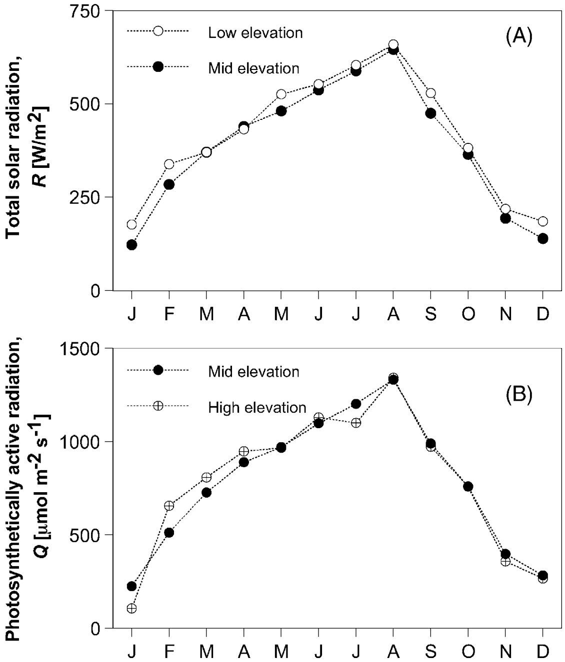

Fig. 10. Monthly mean values for mid-day solar radiation fluxes at different elevations on Mt. Moosilauke. (A) Global radiation (R, W/m^2) measured at low (247 m asl) and mid (748 m asl) elevation and (B) PPFD (Q, μmol m^{-2} s^{-1}) measured at mid and high (1425 m asl) elevation (October 2001–September 2002).

tion) at mid, compared to low, elevation. In the northern Appalachians, Markus et al. (1991) found that elevations between 900 and 1300 m were more likely to experience cloud impaction (typically stratus clouds) than elevations above or below this range, and Miller et al. (1993a,b) determined that forests on Whiteface Mt. were immersed in cloud for 10% of the year at 1050 m but perhaps as much as 35% of the year at 1350 m. In comparison, monthly mean Q (during the growing season, at least) was quite similar between mid and high elevation sites (Fig. 10B), which suggests that any increased frequency of high elevation clouds is offset by the increased Q flux under clear-sky conditions at high elevation. Therefore, these data do not support the hypothesis that mean PPFDs at higher elevations are dramatically reduced compared to those at lower elevations, as has been suggested previously (e.g. Richardson and Berlyn, 2002). In other mountain ranges, previous studies have documented dramatically different situations, and there appears to be considerable variation in the elevation–radiation flux relationship (Körner, 1999). In the upland areas of mid-Wales and the Pennines of Scotland, Harding (1979) noted that monthly sums of solar radiation generally decreased with increasing elevation, except that the summits of the highest peaks (e.g. Scotland's Cairngorm, 1245 m) were often above the clouds and could receive as much as 35% more radiation than nearby lowland areas. In our region of study, cloud immersion frequency appears to differ among mountains (or at least mountain ranges, Mohnen, 1992), and thus the patterns on Mt. Moosilauke may not represent the conditions on, for example, Whiteface Mt. Furthermore, climate change could potentially alter the elevation–radiation flux relationship by shifting the vertical profile of cloud immersion frequency: in our area of study, a 30-year rising trend in the cloud base height has been recently documented (Richardson et al., 2003). This trend could also affect the altitudinal gradient of relative humidity described above, and, perhaps more importantly for montane forests cloud-water deposition to the spruce–fir zone.

4. Concluding remarks

In the mountains of the northeastern United States, as in many mountain regions around the World, vegetation patterns are driven by climate, but the relationships between microclimate and elevation have not yet been adequately characterized. The present study adds considerably to our knowledge not only of the microclimate of montane spruce–fir forests (a globally important forest type) in general, but also the degree to which this microclimate may vary at a regional scale. Two important results have implications for montane regions outside of the limited geographic range we studied.

First, inter-mountain differences in mean annual air temperature at treeline indicate that there are other factors, in addition to temperature, which contribute towards determining treeline location in the mountains of the northeastern United States. More generally, however, this suggests that it can be very difficult to generalize climate–vegetation relationships, even within a region.

Second, these data clearly highlight some of the hazards of extrapolating low elevation data to high elevation sites. For example, although we confirmed a general elevation–mean annual temperature relationship that holds across the area of study, there was, at shorter time scales, significant variation in the lapse rate up the side of each mountain, but the pattern of variation was not consistent among mountains. Also, the altitudinal gradient of soil temperature was less predictable than has been previously reported. And, the altitudinal patterns for solar radiation clearly depend on whether the sky is clear or cloudy, and, in cloudy conditions, the distribution of those clouds along the elevational gradient; the overall effect of elevation is thus uncertain.

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