Honor Paper Award:

Air Temperature Variation Over Snow-Covered Terrain

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ABSTRACT

December 1989 was not only one of the coldest months for which instrumental records exist in the Northeastern United States, but was also unusual in that the air temperature remained continuously below freezing during all but the last day of the month. This prolonged cold period provided relatively homogeneous meteorological conditions in which to study the relationship among complex terrain variables and early morning air temperatures. An experiment was conducted in the Connecticut River Valley near 43°N latitude, based on the hypothesis that the river pool above Wilder Dam would provide a homogeneous surface reference for comparison of air temperatures observed nearby in differing geographic settings. Temperatures were measured 1.5 m above the surface at 92 relocatable points along a 33-km north-south transect and a 12-km east-west transect. Morning twilight temperatures measured on five days prior to a 30-cm snowfall on 16 December were compared with temperatures at the same locations on the five following days. Prior to the snowfall, the temperatures near the river were higher than those immediately upslope by more than 2°C. This trend was reversed following the snowfall, with colder air near the river. An analysis is presented to demonstrate that the heat rejected from river ice growth would be sufficient to provide the observed local warming. The quantity of heat available from this source decreased by an order of magnitude coincident with the observed reversal in temperature trends near the river. The influences of terrain slope, vegetation and the "heat island" of a village are also discussed.

INTRODUCTION

The nature of atmospheric flow, exchange and plume dispersion over complex terrain is of current interest in environmental impact studies and air pollution research (Clements, 1989). Local variations in temperature are of continuing interest in agriculture and in the construction, energy and utility industries, as are rural/urban temperature differences. Terrain-induced and rural/urban temperature differences seem to be enhanced by snow cover, which amplifies the problem of identifying climatic fluctuations in temperature records (Karl et al., 1988; Jones et al., 1989; Landsberg, 1981; Robinson and Kukla, 1988), and complicates long-range weather forecasting (Walsh and Ross, 1988). Most specifically, these small-distance scale variations in near-surface atmospheric conditions prohibit conclusive extrapolation of single-point meteorological observations and interpretation of remotely sensed surface data.

Local variation in near-surface air temperature, initially a function of radiation, is complicated by considerations of air/surface energy exchange and thermal storage (Miller, 1956 a,b). Overnight radiation deficits alter the structure of the boundary layer (Sasamori, 1968; Yu, 1978; Yamada, 1979; Nieustadt, 1980; Arya, 1981; Andre and Marht, 1982), often decoupling the surface from larger scale and more vigorous circulation above, but coupling the near-surface air and cold snow surface (Miller, 1956b). The relatively dense air formed near the surface tends to flow downslope, but this flow may be accelerated or retarded by terrain (Baines, 1979) and vegetation features (Li et al., 1990).

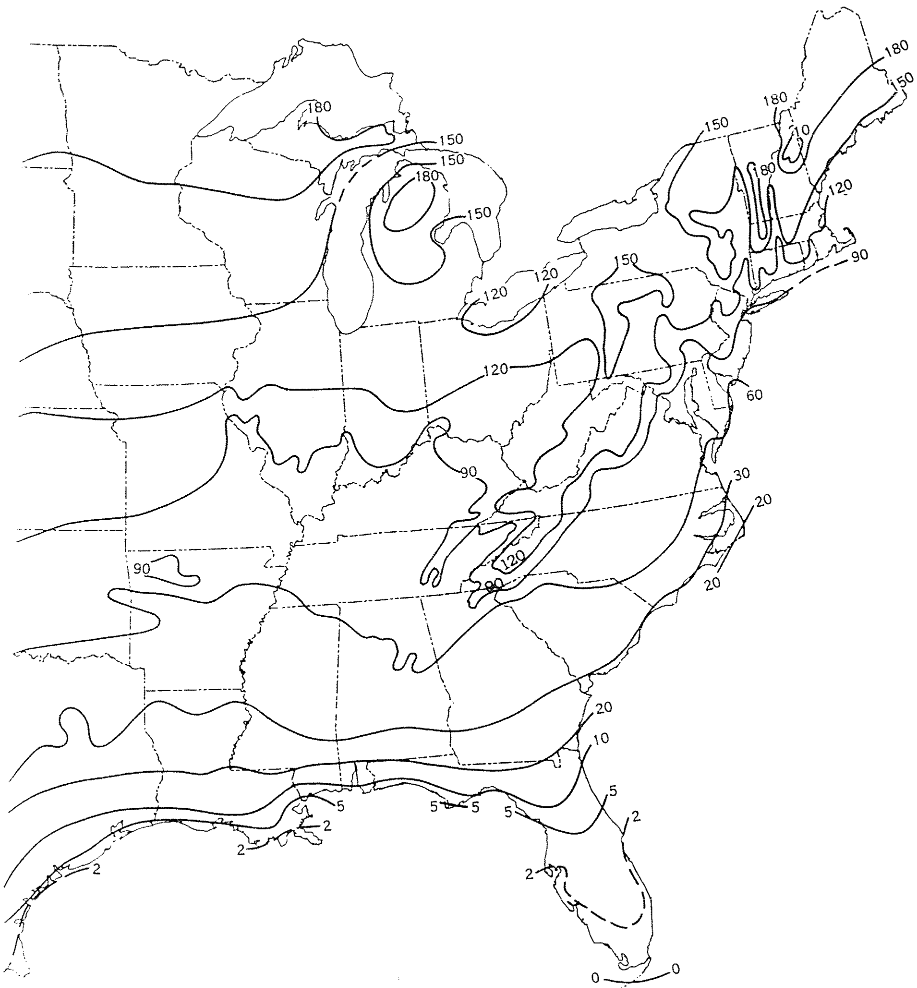

Figure 1. Climatic map of the eastern United States showing the number of days the minimum air temperature is below 0°C [from the National Climatic Atlas, NOAA (1977)].

Maki et al. (1986) and Maki and Harimaya (1988) studied and modeled the heat budget in a basin in volcanic terrain on Hokkaido Island, Japan. The model predicts nocturnal winter cooling to be greater in basins than in flat terrain, and greater at the bases than at the tops of mountains. This is due to the drainage of denser air down slopes and pooling in basins. Mountaintops are not as frequently decoupled from tropospheric flow, as they may remain above the inversion. Air entrained to replace that draining down the mountain may be much warmer than that in the decoupled surface layer below.

The Connecticut River Valley between Vermont and New Hampshire is anecdotally cooler than regions at equal latitude 100 km to the east or west. Later blooming of plants, later and earlier frosts, and absence of some particular types of vegetation native to nearby areas characterize many places in the region. The climatic map in Figure 1 indicates the complexity of the temperature conditions in the northeastern United States. As the observation station spacing is comparable to that of the isotherms in this region, the actual spatial variation of temperatures may not be accurately represented. This paper describes experimental measurements of near-dawn surface air temperatures in the Connecticut River Valley near 43°N as part of an investigation of the influence of nocturnal inversions, drainage winds, basin effects and settlement on local minimum air temperatures.

EXPERIMENT DESIGN

A shielded, relatively rapid response thermistor (YSI 400, response 10 s in air) and its digital readout (repeatable to ± 0.1°C) were adapted for convenient mounting on a car or

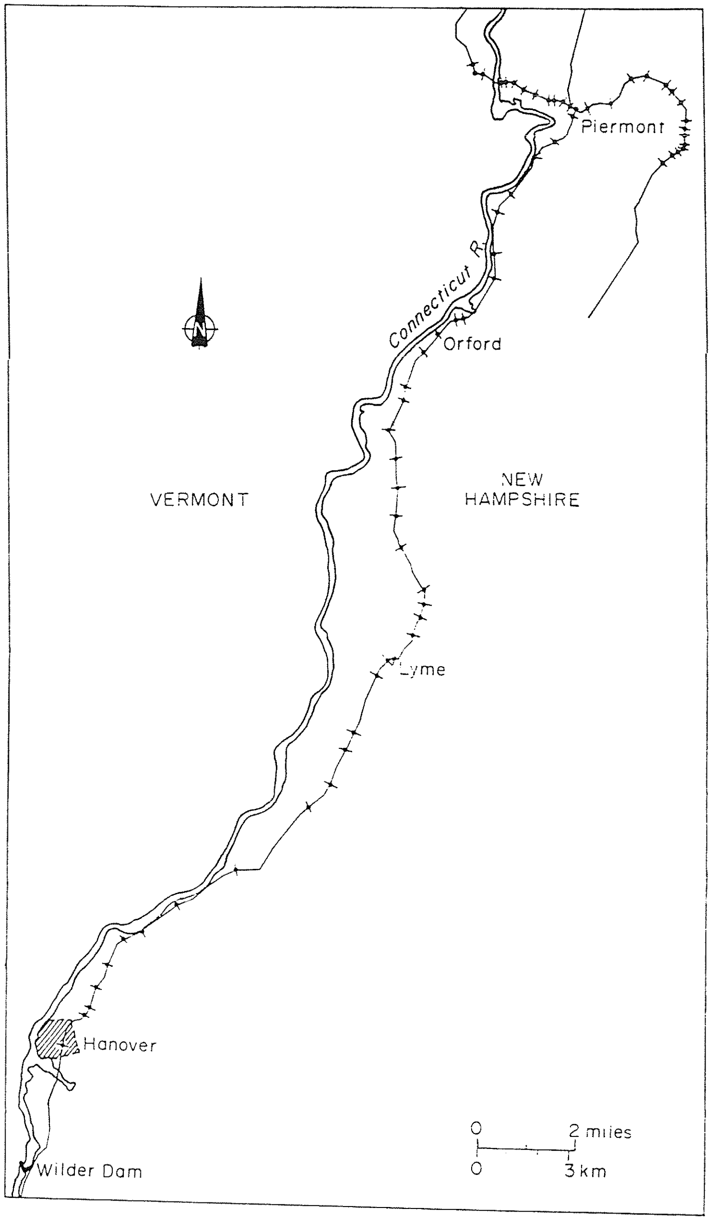

Figure 2. The section of the Connecticut River Valley where the experiment was conducted, including the longitudinal and transverse data transects.

truck. Some experimentation was necessary to select mountings 1.25 to 1.5 m above ground level that were not influenced by radiation or advection of heat from the vehicle. The door-top mount finally selected was found to reflect the free-stream temperature whenever the vehicle was in motion. Experiment and calculation show that the temperature indicated while in motion is representative of an integration of 150 m of horizontal path through the air. This limits the spatial resolution of temperatures to this distance.

Preliminary experiments were attempted during the winters of 1987-88 and 1988-89, and on days when late or early frosts were expected during the spring or fall of those years. The results of these experiments demonstrated that a wide range of temperatures could occur in a 10- by 30-km rectangle with less than 330 m total variation in elevation. A precisely defined measurement protocol was considered essential to subsequent experiments. Observation points which could be precisely located on the USGS topographic maps of the area were selected along an east-west (transverse) transect of the valley at Piermont, N.H., and along a north-south (longitudinal) transect extending from Piermont to Hanover, N.H. A total of 96 observing points (Fig. 2) were defined and noted by landmark on a standardized data

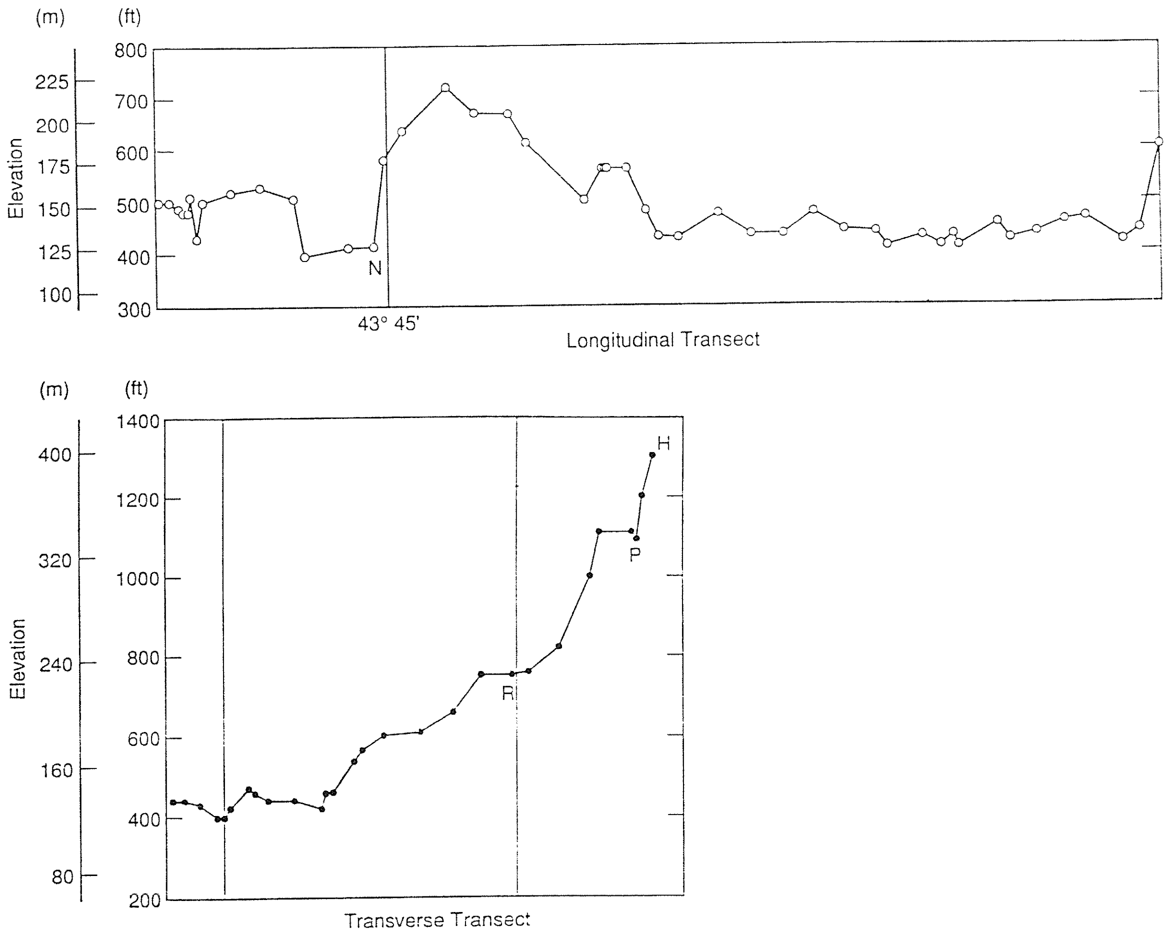

Figure 3. Elevation and location of observation points projected to a north-south line for the longitudinal transect and an east-west line for the transverse transect.

sheet. The analysis was conducted by projecting the terrain cross-sections to east-west and north-south transect lines and entering observed temperatures above the projection. The cross-section projections are shown in Figures 3a and b.

Air temperature measurements were made while in motion between the hours of 0615 and 0730 EST during the period 1 Dec 1989 to 30 Mar 1990. This time was prior to sunrise during most of the period and in the shadow of the hills during March. Only days with relatively uniform sky conditions throughout the experiment area are considered in the analysis. Three days were deleted from the analysis due to fog or precipitation in a fraction of the area. A total of 60 days of observations were obtained that yield some whole winter generalizations. This paper focuses on the results obtained during the relatively homogeneous cold period of December 1989.

We propose two hypotheses as the basis for our analysis of minimum local air temperatures. First, if decoupling of the air in the valley from the lower tropospheric flow is occurring, a discontinuity will be present, and the surface temperature will not be isentropically related to that of the free air above. Second, a conservative plane of reference that is level and homogeneous and extends throughout the experiment area is necessary to compare air temperatures in adjacent environments with variable topography, vegetation or population, within the decoupled layer. The Connecticut River is dammed south of Hanover and presents a level pond throughout the experiment area. The water in this pond is exchanged frequently, preventing stagnation and development of warm or cool pools. We propose that this frequent water exchange will provide a relatively stable heat source or sink to the air at the lowest point in our experiment area and provide a homogeneous reference plane for comparison of temperatures measured in other environments within a reasonable distance

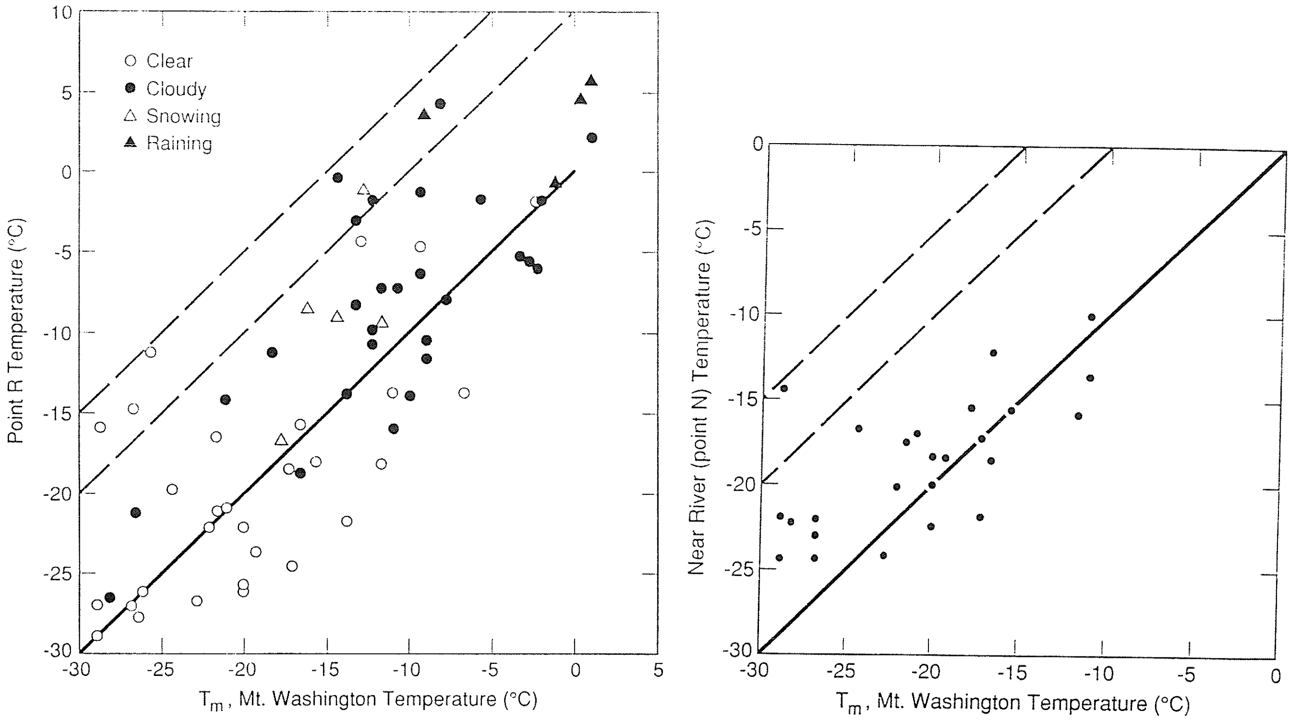

Figure 4. Temperature comparisons of data at points R and N with data recorded at the summit of Mount Washington.

of the river. The river was frozen and snow-covered during most of the period reported, providing an even more stable reference plane.

RESULTS OF EXPERIMENTS

A comparison of temperatures measured at river level, 4 km north of the CRREL recording thermometer (point N), another point in a tributary valley (point R), and the 0700 LCL [12 UT] observation made at the summit of Mt. Washington (80 km east) is shown in Figure 4. The summit of Mt. Washington is nearly 2 km above sea level, and several hundred meters higher than surrounding terrain. The temperature observed there is representative of the lower tropospheric potential temperature over inland New England. The 12 UT observation is synoptic with upper air observations to allow further analysis of the deviation of the mountain temperatures from free-stream temperatures. If no inversion or decoupling were present, adiabatic exchange would provide surface air temperatures in the Connecticut Valley about 20°C higher than Mt. Washington (T_M) in dry air, and 10 to 15°C higher in cloudy air. Figure 4 shows the valley to be always colder than (T_M +15°C), and generally colder than (T_M +10°C). On many occasions the valley temperatures are less than the temperature at the summit of Mt. Washington. Comparison of the temperatures observed at these basin points with nearby hilltop points shown in Figures 5 and 6 indicates the presence of a warmer layer above the valley floor but below the altitude of Mt. Washington. These temperature relations are indicative of one or more temperature inversions between the levels of the experiment points and Mt. Washington. Visual observations showed that near-surface flow was very weak and usually easterly or northerly and occasionally southerly in the valley, while Mt. Washington winds were strong and westerly. The lower mountains between the valley and Mt. Washington apparently interact with the upper winds to produce a weak eddy flow east of the river. The nocturnal inversion decouples the lowest layers from tropospheric circulation. The presence of colder air at the surface than atop Mt. Washington, and a reversal of the near-surface wind direction, supports the hypothesis that the Connecticut Valley is decoupled from the flow above, but definitive proof would require tethersonde or airsonde temperature profiles.

Point N, about 4 km north of the CRREL temperature shelter, was chosen as the basis for comparison of temperatures observed at several points near river level in Dec 1989. The road is immediately adjacent to the river bank for at least 300 m at point N, and a hill immediately east of the road confines circulation, limiting admixture of air from other than

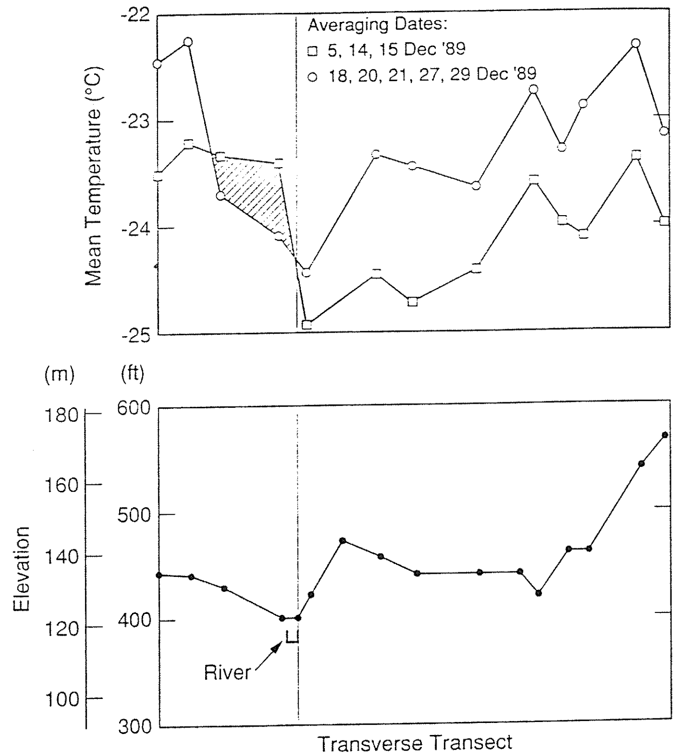

Figure 5. Mean temperatures along the transverse transect near the river before and after the 16 December snowfall.

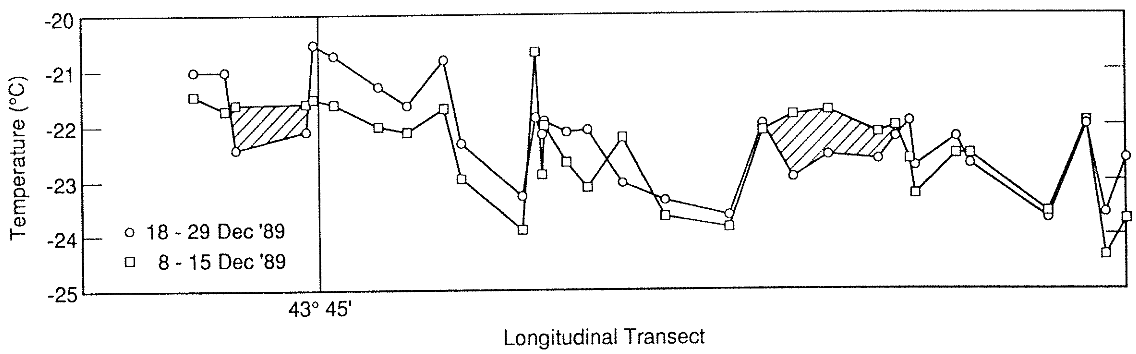

Figure 6. Mean temperatures along the longitudinal transect before and after the 16 December snowfall.

the river direction during still conditions. The temperatures observed at six points near or over the river, displaced in distance by 1.5 to 22 km from point N, are compared with those observed at point N in Figure 7. Examination of Figure 7 shows that 99 of 114 comparison points correspond to within ± 1.5°C; this is considered a tentative verification of the second hypothesis, allowing use of the near-river temperature as reference to examine terrain influence on other nearby temperatures.

Temperatures during December 1989 as observed at the CRREL temperature shelter (about 300 m from the river and 35 m above river level) were less than -15°C on all but four days, and on only one day did the daily maximum exceed freezing. Grease ice was observed on the Connecticut on 1 Dec, and an initial ice cover formed by 4 Dec. A light (5 cm) snowfall on 6 Dec produced a thin snow cover on the river ice throughout the experiment area, which verified that the ice cover was continuous. An additional 25 to 35 cm was deposited in the experimental area on 16-17 Dec. A comparison of temperatures observed prior to 16 Dec and following 17 Dec along north-south and east-west transects of the Connecticut Valley is

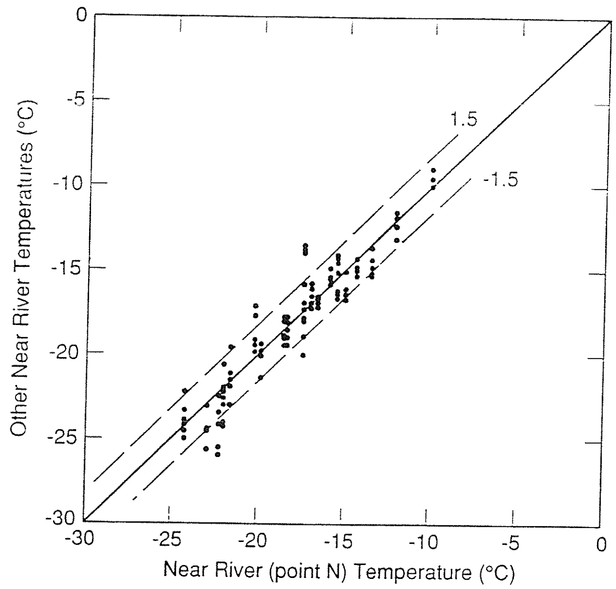

Figure 7. Comparison of temperatures at several locations adjacent to the river with those at point N.

presented in Figures 5 and 6. It is quite apparent that temperatures observed adjacent to or over the river prior to the snowfall are greater than those observed nearby, while temperatures observed at these places following the snow of 16 Dec are less than those observed nearby.

Consistent with our observations we will assume that the river ice growth from zero thickness began on 3 December. Ashton (1989) presented an equation for thin ice growth as

$$\frac{dh}{dt} = \frac{1}{\rho L} \frac{(T_m - T_a)}{\left(\frac{h}{k} + \frac{1}{H_{ia}}\right)} \tag{1}$$

and in integrated form as

$$h = \left[\frac{2k}{\rho L} (T_m - T_a)\, t + \left(\frac{k}{H_{ia}}\right)^2\right]^{1/2} - \frac{k}{H_{ia}} \tag{2}$$

where

- h is ice thickness (m)
- k is thermal conductivity of ice (2.24 W/m °C)
- L is latent heat of fusion (3.34 x 10^5 J/kg)
- ρ is density of ice (917 kg/m^3)
- H_{ia} is the heat transfer coefficient between ice and air (20 W/m^2 °C)
- T_m is the melting point temperature (0°C)
- T_a is air temperature (°C)
- t is time (s)

The temperature data presented in Figure 8 indicate significant warming near the river on 6 December. The heat source is probably the rapid growth of river ice. Therefore, we neglect the insulating effect of the light snowfall in the ice growth calculations. Then, equation (2) yields a mean ice growth of 2.9 cm/day for the period 3-15 December 1989. The Connecticut River has an average width of approximately 150 m in this reach, and ice growth of 2.9 cm/day will produce a daily heat release of 1.33 x 10^9 J/m of river length. Large snowfalls on well-established ice covers provide insulation that resists heat flow and greatly reduces subsequent ice growth. The denominator of equation (1) must then expand to include a term that represents the thermal resistance of the snow layer ($h/k + 1/H_{ia} + h_s/k_s$), where h_s and k_s are thickness and thermal conductivity of the snow, respectively. The thermal conductivity of a snow layer increases with the density of the snow (Ashton, 1986). We will assume that the snowfall of 16 December compacted to a thickness of 15 cm

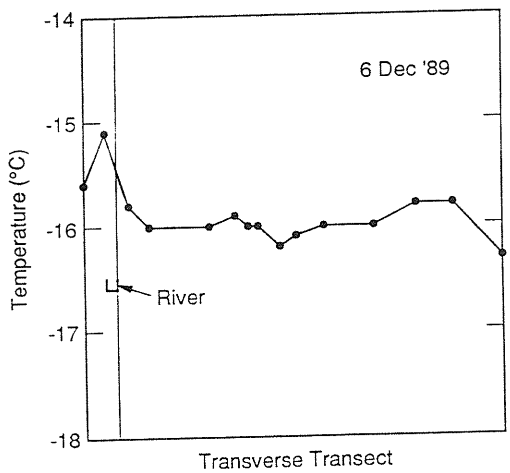

Figure 8. Temperature data from 6 December along the transverse transect near the river concurrent with total freeze-up, verified by continuous snow cover from a light snowfall.

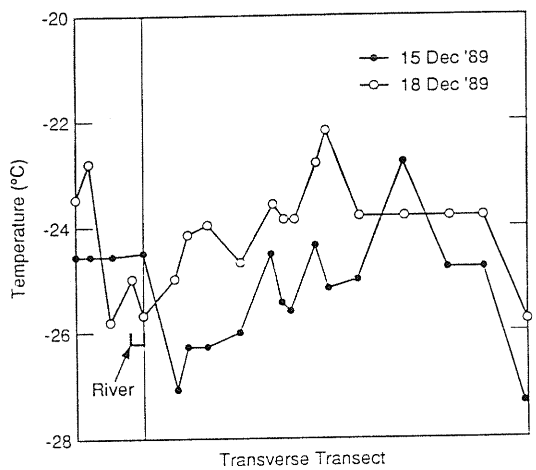

Figure 9. Temperature data along the transverse transect near the river before (15 Dec) and after (18 Dec) the 16 December snowfall.

with a density of 200 kg/m^3 and thermal conductivity 0.12 W/m °C. For an initial ice thickness of 0.38 m, we obtain 0.25 cm of ice growth for 17 December with a corresponding heat release of 1.15 x 10^8 J/m of river length. A comparison with the heat release prior to the snowfall indicates an order of magnitude reduction as a result of the snow cover on the ice.

On 15 December and 18 December the near-river temperatures are approximately equal (Fig. 9). However, the temperature immediately adjacent to the river is about 2°C higher on the 15th, prior to the snowfall. The average temperatures presented in Figures 5 and 6 also indicate relative temperature reductions following snowfall in areas adjacent to the river. We propose that a reduction in local heating resulting from a diminished rate of river ice growth beneath new snow cover was responsible for the observed temperature behavior adjacent to the river. An estimate of the local volume of air that received heat from the river is needed to determine whether differential heat output could cause the observed temperature differences. Observations of smoke plumes from chimneys indicated an inversion at about 25 m above the surface near the river. The temperature data from the valley profiles indicated elevated temperatures, before 16 December, in a band near the river extending through the entire study area with a total width of about 15 river widths. The total volume of air in this band is then about 56,000 m^3/m of river length.

The heat transfer to a system in a constant pressure quasi-equilibrium process corresponds to a change in enthalpy of the volume:

$$ {}_1Q_2 = dH = m\, C_p\, dT \tag{3} $$

where C_p is the specific heat of air at constant pressure, m is the mass of the air and dT is the change in air temperature. At a temperature of -22.5°C and pressure of 1000 mb (100 kPa) the density of air is 1.39 kg/m³ (List, 1950), and the total mass of air in our system is 77800 kg per unit length. The specific heat of air at constant pressure is 1002.3 J/kg K at -23°C (Keenan and Kaye, 1948). Then from equation (3) the quantity of heat required to raise the temperature of the air in the system by 2°C is about 1.56×10^8 J. The difference between the daily heat release from river ice growth prior to and following the snowfall provides about eight times the quantity of heat needed to account for this temperature increase. We conclude that rapid ice growth can cause a significant temperature increase adjacent to a river when this air mass is isolated from large scale exchange. However, a snow cover on the ice can greatly reduce ice production so that the local increase in air temperature becomes negligible.

RESULTS OF EXPERIMENTS: TERRAIN-INDUCED TEMPERATURE VARIATION

Mather and Miller (1967) and Bromwich et al. (1990) have described major terrain-induced air flows over large areas of snow-covered ground in Antarctica. Maki et al. (1986) and Maki and Harimaya (1988) described temperature variation in a smaller basin due to radiation and drainage-induced air flows and showed examples of temperature difference relative to idealized terrain features. Comparing the slope and elevation features of Figure 3 with the temperature features of Figures 5 and 6 shows the importance of terrain considerations in a diverse environment. Slope is a dominant temperature predictor on this scale. Small, flat areas are consistently colder than adjacent slopes; the lowest air temperatures were consistently found above small, flat areas interrupting a slope. Apparently an additional inversion forms above these flats, allowing continuing radiational cooling to occur there, while drainage flow proceeds along the slope and above this additional nocturnal inversion. The flat terrain surrounding point P was just 150 m along the transect, and actual temperatures there may be slightly less than those observed, due to the integration time of the thermistor.

The barriers to flow described by Raynor (1971), Baines (1979) and Li et al. (1990) are present in the form of narrowing or widening of tributary valleys, small ridges, woods and forests. Temperature differences near these barriers were quite pronounced on individual days, but a consistent pattern of significant magnitude is not apparent in the total record. The relative change of slope seems to be the dominant factor which influenced near surface air temperatures during this experiment.

RESULTS OF EXPERIMENTS: THE URBAN HEAT ISLAND

Additional to the temperature differences attributable to terrain, urban/rural temperature differences are well known. Landsberg (1981) showed an immediate effect accompanying the initial development of Columbia, Maryland, when the population was below 10,000 persons. It was immediately apparent in our experiment that the village of Hanover was somewhat warmer than the surrounding areas. The greatest temperature observed on each day was in the downtown area. Perhaps some old New England towns were settled in more climatically benign areas, and are naturally somewhat warmer in winter than the surroundings. Temperatures observed along a golf course near the northern edge of the settled area and those measured between three-story brick buildings on Main Street are plotted in Figure 10 against the temperatures observed adjacent to the river less than 10 minutes earlier. The temperature observed near the golf course is about 1°C less than that observed near the river on half of the coldest mornings when temperatures were less than -15°C; on other occasions the temperatures are generally nearly equal. The temperatures observed in the settled areas are, in all but two cases greater than the nearby river temperature. Twenty-six of the forty-six observations found the settled area more than 1°C warmer than near-river temperatures, and the difference increased as temperature decreased. When temperatures were near freezing, in-town snow removal apparently reduced heat loss, as found in Arctic settlements by Woo and Dubreuil (1983). It appears that a small, non-industrial village can produce a measurable heat island even in complex surroundings.

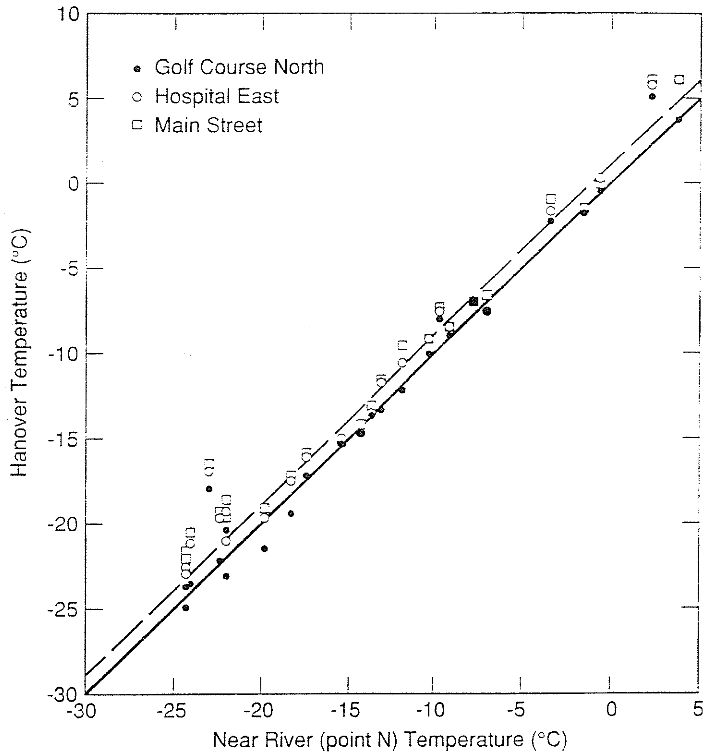

Figure 10. Comparison of temperature data recorded in Hanover with those at point N.

Two additional analyses were performed on hamlets within the experiment area. In these cases, temperatures observed near the center of activity in the settlements and the temperature observed in a sparsely settled flat of similar elevation within a few hundred meters of the hamlet were compared with near-river temperatures. There was no apparent heat island associated with either hamlet, even in the cases when area temperature was less than -20°C.

DISCUSSION

The Connecticut River above Wilder Dam hypothetically provides a temperature reference plane when the adjacent valley is snow-covered. Temperature measurements near and over the river along 30 km of the impoundment, made about the time of sunrise each morning, were used to test and preliminarily verify this hypothesis, when the valley is decoupled from lower tropospheric flow. As the river ice cover is thickening in early winter, the heat released causes the air above and immediately adjacent to the river to be warmed; after the river ice becomes well snow-covered, the air above becomes colder than that nearby in the valley. Although the relative temperature of air over the river when compared to adjacent air varies with ice thickness and snow cover on the ice, the above-river air temperature is quite constant along a 30-km reach, allowing it to serve as a temperature reference plane.

Comparing the temperatures observed in Hanover to nearby over-river air temperatures shows that even a small non-industrial community produces an "urban heat island" on winter nights. Similar comparisons involving hamlets in the vicinity show no systematic temperature difference. This does not eliminate the possibility that heat islands exist at these locations, but indicates that the magnitude is less than that of the terrain-induced temperature variation.

The basin and hilltop variation in temperature presented by Maki and Kikuchi (1986) is apparently amplified over snow-covered ground, due to the decoupling afforded by the multiple inversions present over variably sloped terrain. Relative slope and change in slope appear to produce large changes in early morning air temperature over small distances. Some of these variations may be attributable to vegetation-induced surface roughness and the trapping of air within forests, preventing strong downslope drainage of cold air or promoting eddy circulation of warmer air from above; these variations are not as easily isolated as the heat island and river plane influences and may require a much greater number of measurements to verify or reject.

It appears that the relative coolness of the Connecticut River Valley is due to the decoupling of the air circulation of the valley floor, and of many of the smaller branch basins above, from lower tropospheric flow by combinations of inversions and eddies caused by the north-south axis of the White Mountains. This allows cool pools of dense cold air to form in the valley bottom and in small basins above, and provides "oases" on some hillsides where exchange occurs more freely above local inversions.

Research on this phenomenon is far from complete, and we have only isolated the most apparent terrain/temperature relations thus far. Some practical applications exist, relative to examining the influence of microclimate on historic settlement and abandonment, and perhaps isolating a "rural cold island" that may be generated by extensive cleared fields which are relatively cooler than the antecedent forest.

CONCLUSIONS

A hypothesis has been proposed and preliminarily verified, that a level ice-covered river pool can be used as an air temperature reference when studying elevation-, slope-, and population-induced temperature variation in snow-covered terrain. Changes in the terrain slope and the presence of a non-industrial village both produced significant variations in the local temperature. A research challenge remains, relative to application of this hypothesis to vegetation-induced temperature variation over snow-covered ground.

The presence of colder air, greatly diminished winds and a reversal of the wind direction in a valley relative to conditions aloft has preliminarily verified the hypothesis of a decoupled air mass in the valley during cold, clear early morning conditions with snow-covered ground. The heat released from formation of ice on a river during freeze-up is sufficient to produce 2°C of warming within a band of several river widths in a valley bottom that is decoupled. This heat release diminished to below the level of detection when 15 cm of snow covered the ice.

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