# Investigation of Temperature Variation Over Snow-Covered Ground

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## ABSTRACT

Fragmentary climatic data show that large mean winter temperature differences occur over short horizontal distances in northern New England. Initial winter experiments indicated that very great local variation in pre-sunrise surface air temperature occurred along the Connecticut Valley. A thin layer of air over or adjacent to the Connecticut River was proposed as a reference plane to examine these temperature differences. Experiments showed this reference plane concept to be valid over a 10-km distance scale, but that nonuniform cloud cover often invalidated the reference plane concept on a 30-km distance scale. The influences of slope and terrain on local temperature structure are presented, showing that temperatures on small flats and in small basins differ the most from general tropospheric temperatures. The greatest local air temperature differences with respect to both time and space occur during periods of warm advection, and somewhat lesser variation occurs during cold zonal advection. The least local temperature differences occur during cold meridional advection.

## INTRODUCTION

It is well known that basins and valleys experience significantly lesser overnight low temperatures than nearby ridges or extensive plains (Clements, 1989). Snow cover appears to enhance local minimum temperature differences. This paper describes experimental results from field observations, relative to the time and space variation of air temperature during morning twilight, with respect to snow cover, terrain, and meteorological conditions.

Large-scale katabatic downflow over extensive snow-covered terrain has been examined by Mather and Miller (1967) and modelled by Bromwich et al. (1990). D.H. Miller (1956) extensively studied the influence of snow cover, intercepted snow, and vegetation on air temperature in the Sierras and found that the efficient exchange of heat at the air/snow interface intensified the temperature inversion above, effectively decoupling the surface from the tropospheric circulation. This work combines some of these well established principles of polar and mountain meteorology with smaller distance scale observations of nocturnal inversion formation (Yu, 1978; Yamada, 1979; Nieustadt, 1980; Sutherland, 1980; Arya, 1981; Andre and Mahrt, 1982), exchange across a density interface, valley drainage (Yasuda et al., 1986), and basin flow (Maki et al. 1986; Maki and Harimaya, 1988) to examine the large differences of surface air temperature observed in the Connecticut Valley by Hogan and Ferrick (1990). These small distance scale variations in air temperature influence interpretation of the climatic record, operational forecasting of freezing rain, and dispersion of air pollution.

## METHODS AND INSTRUMENTATION: EXPERIMENT DESIGN

The spatial variation of near-surface air temperature is usually measured by employing a network of recording temperature sensors. This technique has been successfully used by Laughlin and Kalma (1990) to examine frost risk in similarly complex but infrequently snow-covered terrain. The presence of snow cover increases the difficulty of recording near-surface air temperature, and in some cases requires supplementary surrogate measurements to define the air temperature in close proximity to the snow surface (Andreas, 1986). The calm conditions that frequently occur over snow require extensive shielding and aspiration of temperature sensors and also promote the formation of strong local inversions. Aspiration of the sensor can induce sufficient local circulation to modify the ambient air temperature.

The inversion-induced temperature gradient requires frequent visits to the recording station to maintain a constant height of the sensor above the snow surface. This in turn modifies the radiation and conduction pattern in the vicinity of the sensor, altering the near-surface temperature field.

We have chosen to aspirate a temperature sensor by attaching it to a vehicle and moving it rapidly through undisturbed air, to avoid the problems of static temperature measurement. This technique allows a single sensor to be used to measure a large number of points in the brief (1-hr) time window available for these experiments, eliminates the necessity for cross-calibration of instruments, and provides direct measurement of temperature difference without compensation for height or position of the sensor. Experiments and calculations describing the time and distance response of the thermistor sensor used were given in Hogan and Ferrick (1990). The response parameters (Table 1) define the time and space resolution of the temperature measurements reported in this work.

These distances are derived from the manufacturer's time constant, which is 10 s to respond to 0.63 of a step change in still air, and a minimum aspiration speed of 5 m/s to eliminate self heating and the possibility of convective heating by the vehicle. This was experimentally tested by comparing the observed temperature achieved after a 50-m roll at 5 m/s to the nearby free air temperature measured by slinging an identical thermistor. The tabulation indicates that it is possible to examine temperature changes on a less than 100-m spatial scale on untraveled rural roads, which constitute most of the experiment area, and with 200- to 300-m space resolution elsewhere.

**Table 1. Distance required to respond to change in temperature.**

| Vehicle speed (m/s) | Distance to recover (0.95 $\Delta T$) (m) |
|---|---|
| 5 | 60* |
| 8 | 100 |
| 16 | 200 |
| 25 | 310 |

* Minimum aspiration rate.

## RESULTS OF EXPERIMENTS

A series of morning air temperature measurements were begun 6 November 1990 and continued through 16 March 1991. The measurements were made along transects of the Connecticut River valley that were established after analysis of earlier work by Hogan and Ferrick (1990). The east–west transect of the valley was extended eastward to include a slightly higher ridge and a lake basin 200 m above river level. Several observation points were selected along the north–south traverse of the valley. The location and the topography of the experiment area are shown in Figure 1. Although additional

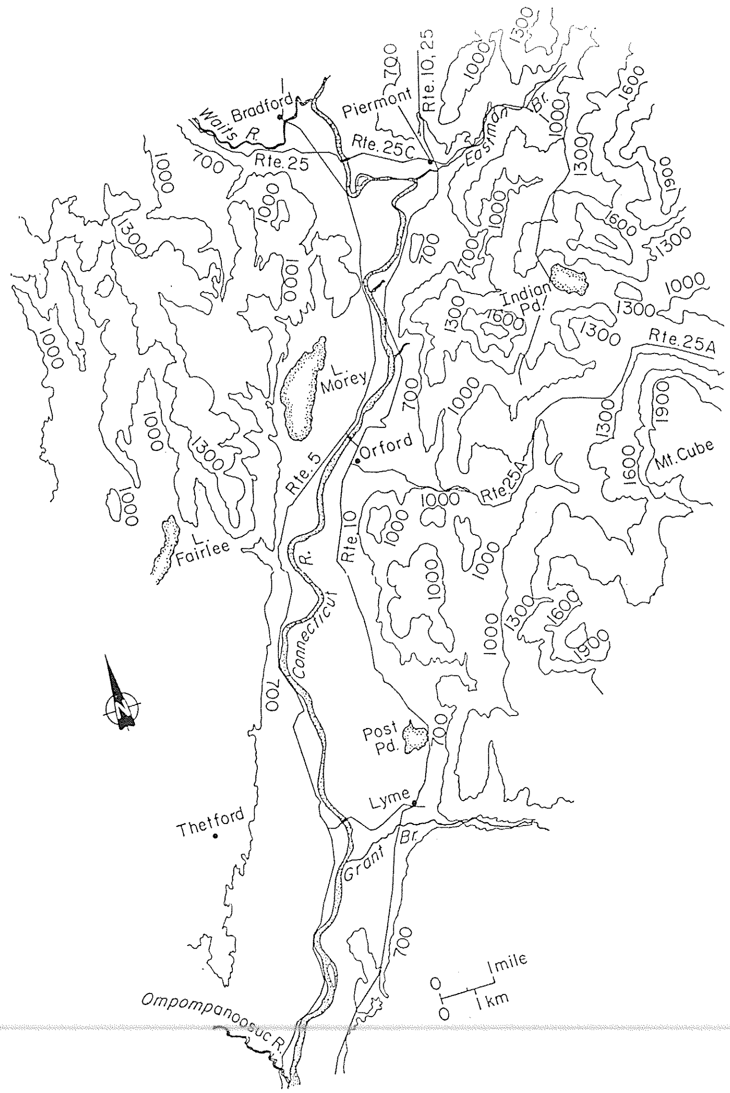


Figure 1. A map of the experiment area along the Connecticut River, showing major topographic features, bridges, and the route of temperature measurement. The contours show approximately 100-m elevation changes, extracted from 300-ft intervals of a USGS quadrangle.

observations are included, the morning observation period remains constrained to 1 hr elapsed time, as in the initial experiment.

The winter of 1990–91 was considerably different in the Connecticut Valley from that of 1989–90. Total snowfall was near the minimum of the century, and falls of 15 cm or more that occurred in November and December were removed by warm advection and rain in the lower elevations along the river within a few days. Sections of the Connecticut River remained unfrozen until late January. A brief core of winter, following 30 cm of snowfall on 13 January and ending in early February, was accompanied by morning temperatures of less than –20°C. An additional complicating factor was non-uniform morning cloudiness, quite unlike recent winters. Orographic clouds formed over the White Mountains and built westward over the river valley on many mornings. On many other mornings, stratiform clouds more typical of the southern sections of the Connecticut Valley were present as far north as Hanover.

The initial hypothesis given by Hogan and Ferrick (1990) proposed that the Connecticut River pond above Wilder Dam would provide a level, homogeneous reference plane that would define the influence of slope, vegetation, culture, and snow cover on early morning temperatures in varying environments within the valley. Preliminary experiments in 1989–90 appeared to support this hypothesis. A dedicated experiment was performed during the winter of 1990–91 to define the space scale over which this hypothesis might be applied. On each morning the temperature was measured at three points (east portal, center, and west portal) of the approximately 100-m-long bridge connecting Piermont, New Hampshire, and Bradford, Vermont. About 3 min later the measurements were repeated in reverse sequence. A coincident pair of measurements was made on flat pasture land of the same elevation about one bridge length east of the bridge.

A slightly shorter bridge connects Orford, New Hampshire, with Fairlee, Vermont, 9 km south of these points. The temperature at the east and west portal of this bridge was observed with similar time separation about 10 min after the observations on the more northern bridge. A comparison of the temperatures measured on the two bridges is provided in Figure 2. The maximum and minimum of the six temperatures observed on each occasion on the Piermont–Bradford bridge define the horizontal dimensions of data boxes; the maximum and minimum of the four temperatures measured 10 min later on the Orford–Fairlee bridge define the vertical dimensions of the boxes. These boxes contain all of the data points observed, and they allow both daily and seasonal variations in temperature difference to be displayed on a single plot. Note that the horizontal and vertical dimensions of the boxes are most frequently quite similar and are constrained by ±1°C of absolute temperature difference. Two systematic departures occur in the 67 data pairs.

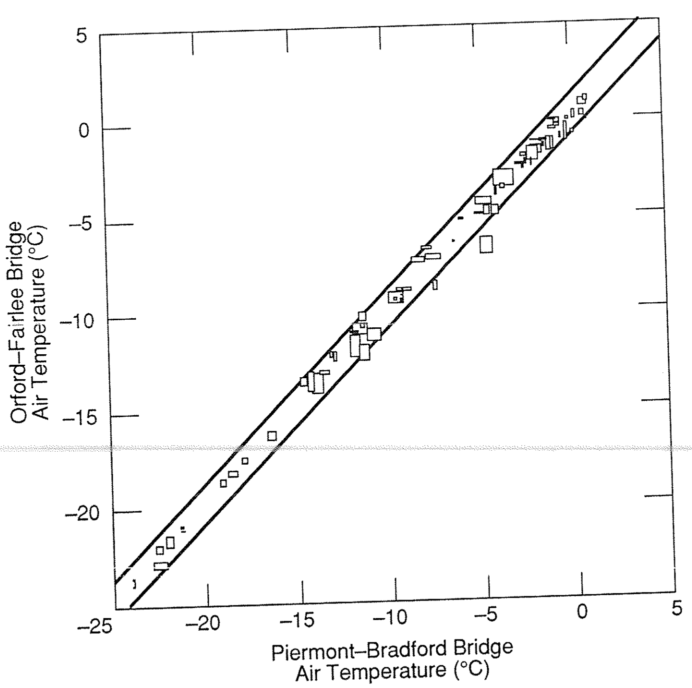


Figure 2. A comparison among surface air temperatures measured at 6 points along the Piermont–Bradford bridge and 4 points along the Orford–Fairlee bridge on 67 winter mornings. The range of temperatures measured is enclosed by the plotted box in each case.

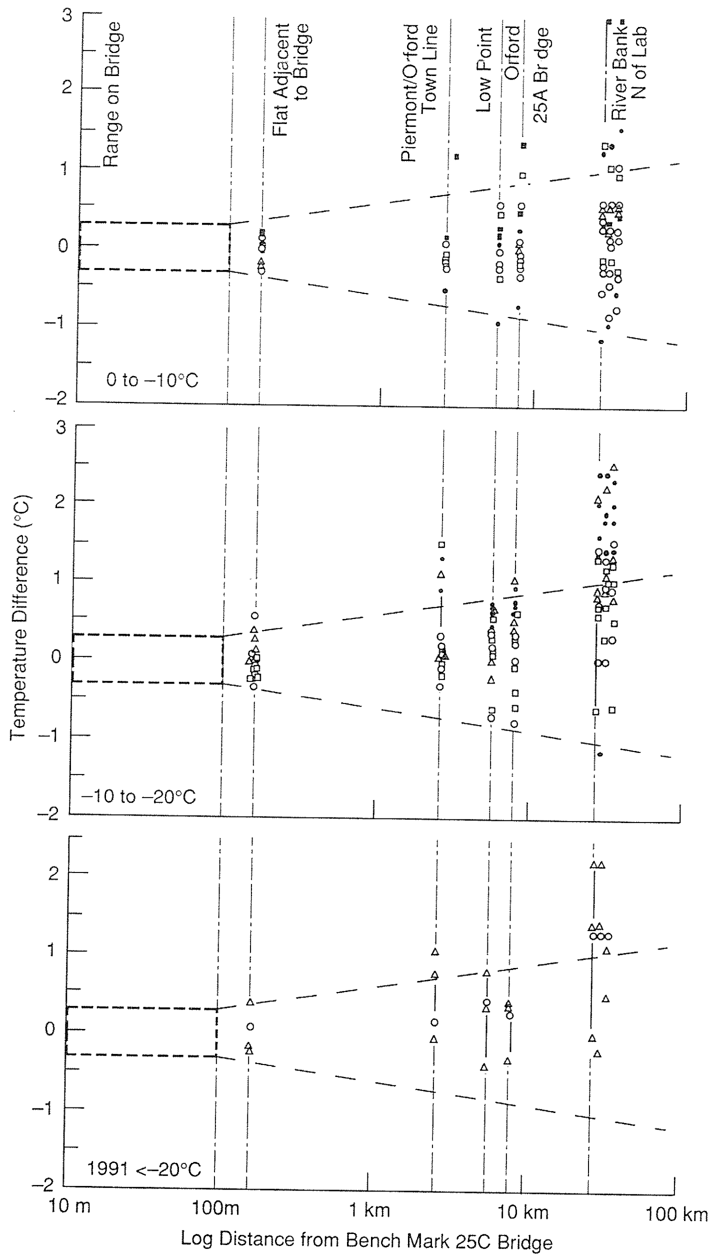


Figure 3. Early morning air temperature difference measured over or along the Connecticut River during winter 1990–91, stratified by ambient temperature classes. The distance scale originates at a benchmark near the west portal of the Piermont–Bradford bridge, and all distances are along radial lines connecting the measurement point with the benchmark. The plotted box corresponding to the length of the bridge indicates the maximum difference in temperature measured along the bridge; the lines extending from that box are the envelope of temperature variation calculated from eq 1. The temperature differences plotted were obtained by subtracting the temperature measured at that point from the mean of the 6 temperatures measured on the bridge.

In addition to measurements made on the flat 100 m east of the Piermont–Bradford bridge, the temperature was observed at four similar points within 100 m of the river and at 5 to 10 m greater elevation than the river. The temperatures observed at these points on mornings when the ground was snow-covered are stratified by ambient temperature; they are plotted in Figure 3 as a function of the logarithm of distance from a survey mark on the west end of the Piermont–Bradford bridge. The maximum difference among the six points measured on this bridge is enclosed by a plotted box corresponding to the relative location of the bridge. There was less than 0.1°C temperature variation along the bridge on only six occasions; a maximum variation of 0.6°C was observed, and the most frequent variation was 0.3°C. The temperature observed on the flat one bridge length from the bridge varied from the mean temperature measured on the bridge by less than the range of variation of the on-bridge temperatures on most occasions. The most frequent variation is ±0.2°C (with respect to the mean temperature measured on the bridge), which is the minimum detectable variation. This similarity among temperatures measured along the river and over the river indicates that the near-river measurements are representative of the river reference plane temperature.

The three panels of Figure 3 stratify the temperatures measured by gross meteorological characteristics, using temperature range as a classification criterion. Mornings with temperatures of from 0 to –10°C are generally overcast; "typical" Connecticut Valley winter mornings have temperatures of –10 to –20°C, and clear, calm mornings are usually associated with air temperatures colder than –20°C, although cirrus may be present on some cold days. The difference in temperature [$\Delta T_d$] measured at the several points with respect to distance [$D$] can be related to the temperature difference [$\Delta T_b$] over the 100-m length [$L$] of the Piermont–Bradford bridge through an empirical relation:

$$[\Delta T_d] = [1 + \log(D/L)]\bullet[\Delta T_b] \tag{1}$$

where $D$ is the distance from the reference at the west portal of the bridge to the location of the second temperature observation. This formula was not physically derived but empirically generated through dimensional analysis (Stull, 1988) and consideration of other variations in temperature with height and distance that are logarithmically related. It is used here to constrain the envelope of temperatures observed along the river reference plane, for comparison and analysis.

The data observed on the cloudy (0 to –10°C) and clear, calm (< –20°C) mornings of 1990–91 are constrained quite well to the envelope defined by this empirical formula. Examination of the central panel, displaying mornings with temperatures of –10 to –20°C, shows that few days are colder at the riverside near the laboratory, 30 km from the reference point. About one-third of the days have greater temperatures than the constraining band. Two north–south traverses of the valley, obtained on 11 and 14 Jan 1991, are shown in Figure 4. The traverse of 11 Jan represents a morning in the –10 to –20°C range, and the traverse of 14 Jan provided the maximum positive temperature difference in the lower panel of Figure 3. Although the reference temperature on 14 Jan was –22°C, it is a day more typical of warmer mornings, as the sky was more than one-half covered by relatively thick cirrostratus [Cs] clouds. Comparison of Figures 3 and 4 indicates that when partial cirrus cover overlays snow-covered ground, a systematically greater air temperature occurs in the southern sector of this segment of the Connecticut Valley than in the northern sector. This

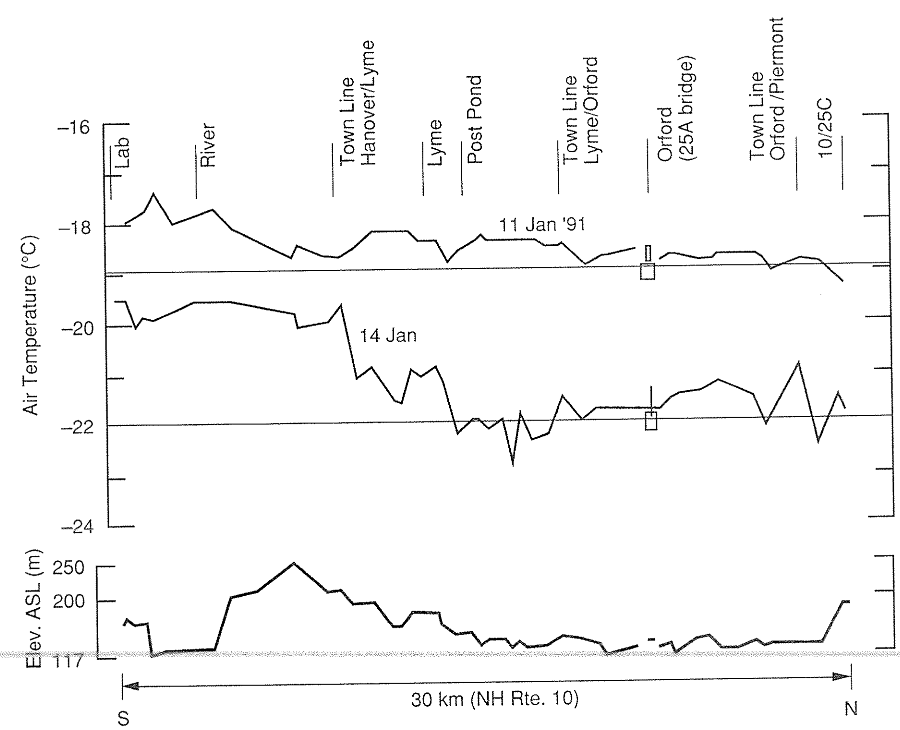


Figure 4. Early morning air temperatures measured along the Connecticut River Valley on two partly cloudy mornings. The elevation cross section is shown at the bottom, referenced to the river surface at 117 m. The points shown in the left third of the figure are systematically warmer and exceed the envelope calculated from eq 1, but the northern points are well constrained by the envelope.

is in contrast to the relatively uniform temperatures through this 30-km extent that were observed during the generally clear mornings of 1989–90.

This experiment is constrained by a narrow time window that is bounded by sufficient twilight to recognize reference landmarks at the beginning of each morning's experiment and a time about 15 min after astronomical sunrise. This provides about 1 hr to visit each observation point sequentially. It is possible that some points, under some meteorological conditions, may experience temperature variations comparable to the station-to-station differences used in data analysis during the experiment period. The difficulties in measuring temperature over snow essentially preclude using a recording thermograph to monitor for short term variability. The nature of the station measurement route does facilitate replicate measurements of individual points, at time intervals of 2 to 15 min. In addition, the multipoint, multitime measurements made on the two bridges provide a minimum variation amplitude applicable to the measurement scheme.

The routine measurement protocol used in 1990–91 included replicate measurements at 33 of 38 east–west transect stations at 3- to 15-min intervals. On two occasions, the southernmost leg of the north–south traverse was eliminated and complete replicate sampling of the east–west transect and the northern portion of the north–south traverse was completed in 1 hr. One of these (0629–0723 EST, 26 Jan 1991) coincided with the coldest air observed during 1991; it is shown in Figure 5. The observed air temperatures are superimposed on an elevation cross section of the east–west transect route across the Connecticut Valley. The transect is projected to a straight line connecting the extreme east and west stations, and interstation distance is conserved. Landmarks are provided to allow orientation of the transect route relative to Figure 1. The Connecticut River pond is the lowest elevation at 117 m ASL; large flats lie along the

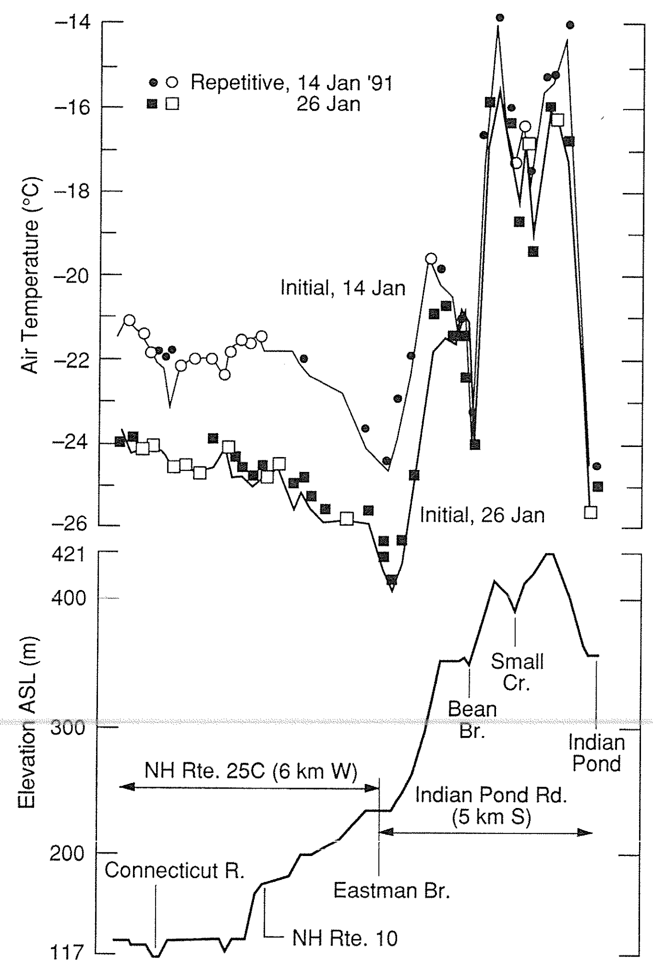


Figure 5. Temperatures observed along an east–west transect of the Connecticut Valley on two mornings when warm advection was occurring. The initial temperatures measured are connected by plotted lines. Coincident repetitive measurements are indicated by open symbols, and noncoincident repetitive measurements are indicated by solid symbols. The elevation cross section is shown at the bottom.

river at 135 m ASL. Small flats or fairly level basins lie along the route at 230 and 335 m ASL, and the transect terminates in a lake basin at 340 m ASL. The highest point along the transect is at 421 m ASL, but several hills and ridges of 1000 m ASL are within a few kilometers.

The transect of 26 Jan began at the 230-m basin at 0629 and proceeded east over the highest elevations, reaching the lake basin at 0642 and returning to the 230-m basin at 0657. The west leg then commenced immediately, reaching the west side of the river at 0709, and returning to the 230-m basin at 0717. Initial air temperatures are connected by the plotted lines: repetitive temperatures are noted by open symbols on the plotted line if less than 0.1°C of difference occurred. Solid symbols denote repetitive temperatures differing by more than 0.1°C. Temperature differences are on the order of 0.5°C or less in all but four cases, and 14 of 45 stations varied by less than 0.1°C. The instances where temperature differences of 1.0°C were found were adjacent to changes in slope and apparently related to intersecting layers that varied greatly in temperature over small vertical distances. A similar illustration is given in the second trace of Figure 5 to verify the consistence of this behavior.

The temperature differences observed at each point on 26 Jan are plotted as as a function of distance and time in the lower panel of Figure 6. Temperature differences observed on the calm, clear morning of 8 Jan, during a period of cold advection, and the morning of 11 Jan, which was solidly overcast but also a period of cold advection, are plotted in the upper panels. It is apparent from Figure 6 that temperature variation with respect to time periods of less than 30 min is small on days characterized by cold advection, but relatively great over the same time periods on days of warm advection. Ridge tops and slopes display the greatest time variation of air temperature during warm advection, but flats, v-valleys, and basins in close space proximity to these ridges exhibit very little variation in surface air temperature during the same time periods.

A feature of measuring near-surface air temperature through aspirating the sensor by advancing it through undisturbed air is that such a moving sensor may penetrate inversions that intersect slopes. Analysis of observations with respect to elevation of the station provides the vertical structure of the boundary layer in complex terrain, and deviation of the temperature at a station from that of other stations at the same elevation provides insight relative to the influence of terrain features on local temperature.

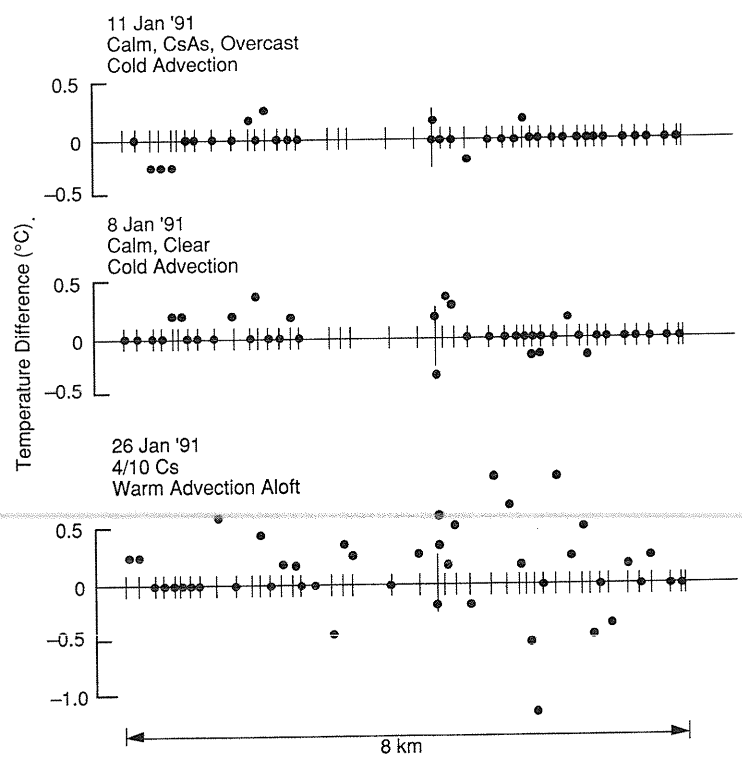


Figure 6. The range of temperatures measured at each point along the east–west transect, on two mornings when cold advection was occurring, are shown in the upper panels. The range of temperatures observed 26 Jan under warm advection (from Fig. 5) is shown in the lower panel for comparison. Temperature changes of less than ±0.1°C over the period of 3 to 12 min between replicate temperature measurements are plotted as zero change.

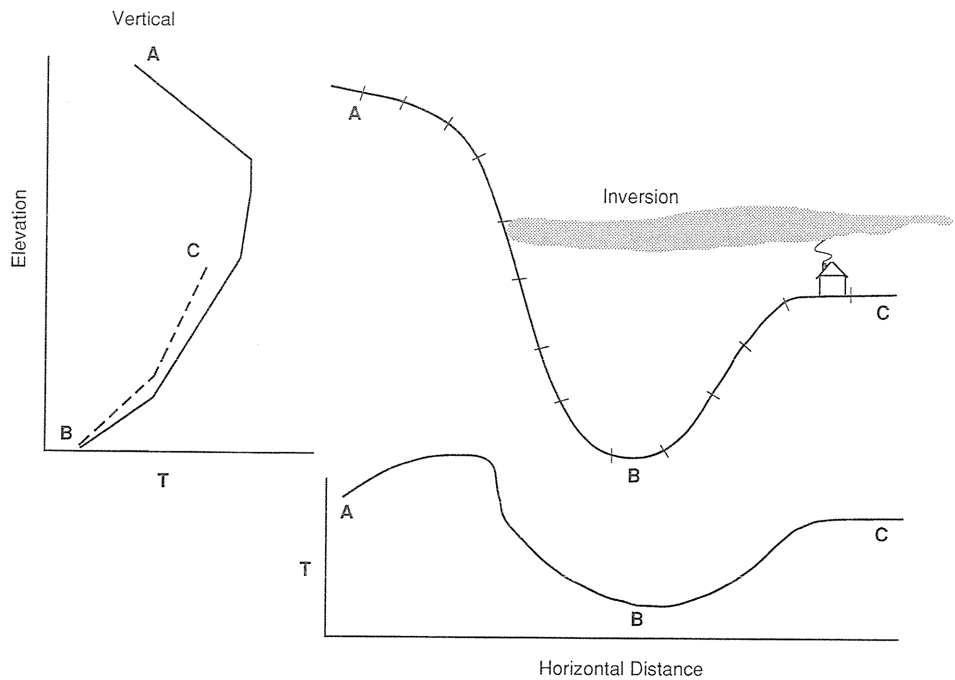


Figure 7. A diagram of the technique used to extract the vertical temperature structure from surface observations along a valley transect.

A diagram illustrating the technique for analyzing horizontal and vertical gradients is shown in Figure 7 for a cross section of a small basin. The smoke from the house chimney in the diagram flattens at the inversion and spreads. A series of temperature measurements along a transect of the basin ABC would produce the spatial variation in temperature shown below the cross section. Plotting the temperatures as a function of station elevation would yield the vertical profiles AB and BC shown to the left of the diagram. This presentation greatly aids the understanding of finding the greatest temperatures near brows and along slopes during inversion conditions and under warm advection.

The east–west traverse of 14 Jan 1991 shown in Figure 5 is translated to a vertical cross section in Figure 8. Analysis of this vertical temperature profile indicates that a very strong inversion of 40°C/km is present above about 230 m. Below this elevation a lapse condition of near or perhaps even exceeding the adiabatic rate of –10°C/km seems to be present through comparison with the plotted adiabat. This lapse condition may be related to ice-crystal-induced (Gotaas and Benson, 1964) or other radiational cooling around the 250-m level, but requires systematic study. The temperatures observed along the flats and basins indicated in the elevation cross section of Figure 5 lay to the left (colder) of the plotted inversion and lapse rate. Temperatures measured along slopes lie to the right (warmer) of the plotted profiles and are in general the points that display the maximum variation in temperature with respect to time in Figure 6. The several values of temperature present at a single elevation along these several flats and basins is indicative of an additional low inversion decoupling the flat from the warm air above.

## DISCUSSION AND CONCLUSION

The winter of 1990–91 was not ideal for examining temperature variation over snow-covered complex terrain but it did indeed advance an understanding of a meteorological regime with extreme spatial variability. The stratiform cloudiness often observed over the Connecticut Valley to the south of Hanover may have been an indicator of a meridional flow of warmer air more deeply into the valley, contributing to the generally mild winter. This forced a relaxation of the uniform cloud cover criteria for examining spatial temperature variation, but produced a more generally applicable technique for estimating the temperature variation along the river surface reference plane. We propose to expand observations to the north and south along the Wilder pool to examine the general applicability of eq 1 in predicting the envelope of winter temperature variation in a basin and to evaluate the possibility of flow divides aligned with identifiable terrain features.

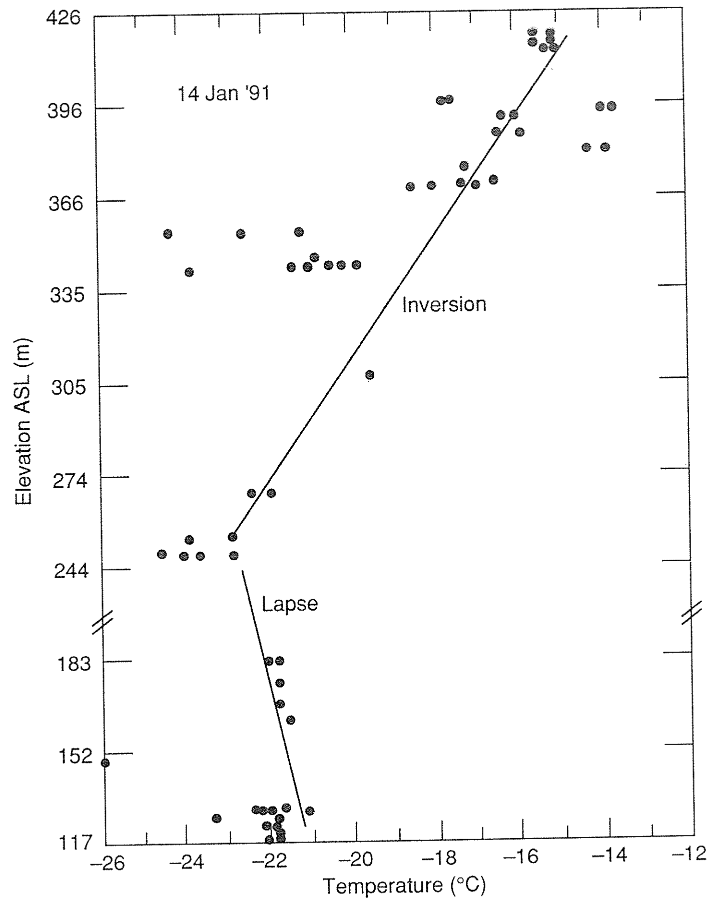


Figure 8. Vertical structure of the air temperature in the Connecticut Valley on 14 Jan 1991, extracted using the technique diagrammed in Figure 7. The plotted line extending from 117- to 230-m elevation is a dry adiabat; the plotted line above 230 m is an inversion of 40°C/km. Temperatures measured along slopes (as shown in Fig. 5) lie near the plotted lines. Flats and basins influenced by near-surface inversions lie to the left (cold side) of the plotted lines; ridges exchange more readily with the warm lower troposphere and lie to the right (warm side) of the plotted lines.

Morning surface temperature variation on a 5- to 20-min time scale appears to be a good indicator of the vigor of local exchange during warm advection. Very little temperature variation was found on this time scale in flats and basins, when variations of a degree or more were present in nearby areas exchanging with warmer tropospheric flow over the same intervals. Very little temperature variation was found on the 5- to 20-min time scale at any station during periods of cold advection. Advection of cold dense air rapidly results in exchange to the surface in even this complex terrain, and no decoupling was evident, even on those clear days when radiational cooling might have been expected to be quite rapid.

The inversion strength of 40°C/km observed across the Connecticut Valley approaches that of the polar regions. It is quite effective in delaying the advection of warm air to the surface and contributes to the diminution of the mean winter temperature of the valley with respect to other North American regions of similar latitude. Our experiments to date indicate this decoupling of the surface from the troposphere above is most effective during warm advection and cold zonal advection over snow-covered ground. We propose that the heat exchange to the snow surface quickly reduces the temperature of the surface air, producing additional inversions that enhance the cooling feedback mechanism.

## BIBLIOGRAPHY

Andre, J.C. and L. Marht. 1982. The nocturnal surface inversion and influence of clear air radiative cooling. *J. Atmos. Sci.*, 39: 864–878.

Andreas, E.L. 1986. A new method of measuring the snow surface temperature. *Cold Reg. Sci. Tech.*, 12: 139–156.

Arya, S.P.S. 1981. Parameterizing the height of the stable atmospheric boundary layer. *J. Appl. Meteorol.*, 20: 1192–1202.

Bromwich, D.H., T.R. Parish and C.A. Zorman. 1990. The confluence zone of the intense katabatic winds at Terra Nova

Bay, Antarctica, as derived from airborne sastrugi surveys and mesoscale numerical modeling. *J. Geophys. Res.*, 95: 5495–5509.

Clements, W.E. 1989. Organizer, Atmospheric Studies in Complex Terrain: Part I and Part II. *J. Appl. Meteorol.*, 28: 405–542, 545–689.

Hogan, A.W. and Ferrick, M. 1990. Air temperature variation over snow-covered terrain. *Proc. East. Snow Conf.*, 47: 1–12.

Laughlin, G.P. and J.D. Kalma. 1990. Frost risk mapping for landscape planning: A methodology. *Theor. Appl. Climatol.*, 42: 41–51.

Maki, M. and T. Harimaya. 1988. The effect of advection and accumulation of downslope cold air on nocturnal cooling of basins. *J. Met. Soc. Japan*, 66: 581–597.

Maki, M., T. Harimaya and K. Kikuchi. 1986. Heat budget studies on nocturnal cooling in a basin. *J. Met. Soc. Japan*, 64: 727–741.

Mather, K.B. and G.S. Miller. 1967. Notes on the topographic factors affecting the surface wind in Antarctica, with special reference to katabatic winds and bibliography. REP UAGR 189, 125 pp. Geophys. Inst., University of Alaska, Fairbanks, 1967.

Miller, D.H. 1956. The influence of an open pine forest on daytime temperature in the Sierra Nevada. *Geog. Rev.*, 46: 209–218.

Nieustadt, F.T.M. 1980. A rate equation for the inversion height in a nocturnal boundary layer. *J. Appl. Meteorol.*, 19: 1445–1447.

Stull, R.B. 1988. *An Introduction to Boundary Layer Meteorology.* Chapter 9. Similarity Theory. Kluwer Academic Publishers, Boston 1988, pp. 347–399.

Sutherland, R.A. 1980. A short-range objective: Nocturnal temperature forecasting model. *J. Appl. Meteorol.*, 19: 247-255.

Yamada, T. 1979. Prediction of the nocturnal surface inversion height. *J. Appl. Meteorol.*, 18: 526–531.

Yasuda, N., J. Kondo and T. Sato. 1986. Drainage flow observed in a v-shaped valley. *J. Met. Soc. Japan*, 64: 283–301.

Yu, T-W. 1978. Determining height of the nocturnal boundary layer. *J. Appl. Meteorol.*, 17: 28–33.