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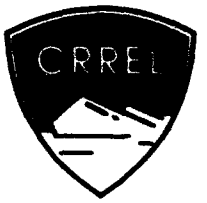

CRREL REPORT 93-19

On the Temperature Distribution Near a Cold Surface

Yin-Chao Yen

October 1993

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Abstract

Temperature profiles were taken during the 1991–92 season at an experimental site on CRREL's grounds; however, they were rather limited because of the unusual lack of snowfall and the location of the site, which was small and had elevated sides (especially in the direction of the prevailing wind). Nevertheless, some unusual features have been observed for the first time. The profiles show the existence of a double-reversal in temperature structure close to the snow surface when the snow temperature is much lower than its melting point, in contrast to a persistent raised maximum temperature over a melting snowpack. This indicates that the micro-heat transport process is complicated by the presence of a nonisothermal lower boundary. Over frozen or partially frozen ground without snow, and with air temperature either above or below the melting point of ice, a thin, nearly isothermal air layer on the order of 2 to 3 cm in thickness at varying heights was observed for the first time. With very few exceptions, in general, the temperature decreases as height increases, but the extent is less pronounced over a snow cover (where conditions are nearly isothermal) than over cold ground. On the basis of limited measurements, the overall temperature gradient over the bare ground is about twice that over snow-covered ground.

Cover: Temperature variations with height over a snow surface as a function of time.

For conversion of SI metric units to U.S./British customary units of measurement consult ASTM Standard E380-89a, *Standard Practice for Use of the International System of Units,* published by the American Society for Testing and Materials, 1916 Race St., Philadelphia, Pa. 19103.

CRREL Report 93-19

U.S. Army Corps of Engineers
Cold Regions Research & Engineering Laboratory

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PREFACE

This report was prepared by Dr. Yin-Chao Yen, Research Physical Scientist, Geophysical Sciences Branch, Research Division, U.S. Army Cold Regions Research and Engineering Laboratory. Funding for this work was provided by the Office of the Chief of Engineers through DA Project 4A161102AT24, *Research in Snow, Ice and Frozen Ground*, Work Unit AT24-SS-E09, *Surface–Air Boundary Transfer Processes*, and DA Project 4A1623734DT08, *Combat Engineering Systems*.

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NOMENCLATURE

a	exponent in the ratio of $(p/p_0)^a$
A	amplitude
b	constant defined as 0.03 R_W, also $(\pi/[24(K_R + K_E)3600])^{1/2}$
B	Planck function
c_p	specific heat at constant pressure, also heat capacity of air
c_v	specific heat at constant volume
E	radiative energy
e	vapor pressure
F	effective flux of longwave radiation
F_s	solar flux
g	gravitational acceleration
h	apparent height of source above its true height
H	hour angle
k	product of $E/\partial T$ and bT/p_w
K_E	$\Sigma m(p_0 - p)/g\rho^2$
k_i	absorption constants for water vapor at wavebands i
K_R	$bT/\rho c_p p_w$
ℓ	height of an air column containing 0.3 mm of precipitable water (= $0.03/\rho_w$ cm)
n	index of refraction for air
p	atmospheric pressure at height z
p_0	atmospheric pressure at $z = 0$.
p_T	arbitrary pressure at the top of the atmosphere
p_w	water vapor pressure
q	specific humidity at height z
q_0	specific humidity at $z = 0$
$q_{r,n}$	upward net radiative flux
R	gas constant
R_W	gas constant for water vapor
S_0	solar constant (~900 W/m^2)
T	absolute air temperature
T_0	absolute temperature at $z = 0$
ΔT_R	temperature change due to longwave radiation flux
ΔT_T	temperature change due to turbulent exchange
ΔT_{T+R}	temperature change due to turbulent and radiative flux divergence
t	time
w	contents of the absorbing substance
x	horizontal distance of the source
x'	range indicated by the instrument
$\left(\frac{\partial T}{\partial t}\right)_{rad}$	rate of temperature change due to radiation
$\left(\frac{\partial T}{\partial t}\right)_{obs}$	observed rate of temperature change
$\frac{dp}{dw}$	derivative of the transmission function
$\hat{Z}$	solar zenith angle
z, Z	vertical coordinate or height above the surface

Greek

ε	$-\frac{\partial F}{\partial z}$ flux divergence
τ	transmission function
ω	precipitable water vapor, also rate of temperature change
γ	c_p/c_v
δ	solar declination
ϕ	latitude
β	absorptivity of the column directly above the level in question, also lapse rate
ρ	air density
α_{da}	dry adiabatic lapse rage defined as $g\gamma/R$
ζ	dimensionless parameter defined as $z^2/4(K_R + K_E)\,t$
$\Psi(\zeta)$	defined as $[(1+2\,\zeta^2)(1\text{-erf}\zeta) - 2\pi^{1/2}\zeta_e{}^{-\zeta}]$
$\phi(\zeta)$	defined as $(1\text{-erf}\zeta)$

On the Temperature Distribution Near a Cold Surface

YIN-CHAO YEN

INTRODUCTION

The air temperature at a couple of meters above the Earth's surface is usually well known because meteorologists and climatologists frequently observe and record it to collect historical weather data and to predict weather. However, the temperature of the ground and especially the snow-covered ground is measured less often, and the temperature distribution within the lowest meter is known largely by inference. To date, very few observations of this distribution have been made, and the reported data usually vary from one researcher to another owing to different interpretations of the results. Added to this problem are the use of probes of questionable accuracy placed in a region having a large temperature gradient and the lack of understanding of the near-ground physical processes that control the temperature.

Suppose there is an equilibrium air temperature resulting from the combined heat transfer processes of radiation and turbulent convection as well as molecular conduction. Consider the case of frozen ground or snow-covered ground (in the latter case, the maximum temperature will be the melting point, i.e., 0°C), which is cold relative to the overlying air. Heat or energy will be transferred towards the cold surface by predominating turbulent convection and less significant molecular conduction, resulting in a continuous temperature distribution with its gradient decreasing monotonically away from the surface. The radiative transfer is hypothesized to alter the convection-established temperature profile in the following manner: The air adjacent to the cold surface exchanges radiation with the surface without much absorption or emission, but it absorbs more radiation from the overlying warm air than it emits. Therefore, the air just above the surface increases in temperature until the increased convective energy transfer equals the net radiation absorbed. The warm air further away from the surface exchanges radiative energy with the overlying air without much net absorption, but it radiates more energy downward than it absorbs from the cold surface. In this way, its temperature decreases until an equilibrium is attained between the radiative and convective processes.

The essential purpose of this investigation is to measure directly the temperature distribution within a distance of about 1.3 m above cold ground (may be frozen) as well as above snow-covered ground and to verify the reported temperature inversion over a snow surface. Subsequently, the commonly used technique—i.e., use of a constant-flux layer in computing the sensible heat flux for energy balance models—will be justified.

REVIEW OF PREVIOUS WORKS

Gaevskaya et al. (1962) reported an analytical study of radiative heat flux divergence and the heat regime in the lowest layer of the atmosphere. With the use of nonstationary heat exchange, they established the importance of the different factors affecting the thermal regime of the near-ground layer and concluded that, in creating temperature change, the combined effects of radiation and turbulence are not additive and that radiative flux divergence has a great influence.

In general, researchers have felt that the turbulent heat exchange is the dominating factor in the heat regime of the lowest atmospheric layer, and that the role of the divergence of longwave radiation flux is negligibly small. They have arrived at this conclusion because of the satisfactory coincidence of the results of observations and theoretical calculations of air temperature, based on the heat conduction equation, that consider

only the turbulent heat exchange, and do not incorporate the influence of the thermal radiative flux divergence on the coefficient of turbulent mixing.

On the basis of theoretical calculations by Kondrat'yev (1969), we find that the magnitude of longwave radiative flux divergence reaches very large values in the lowest atmospheric layer and that the rate of cooling of the air by radiation is much greater than the cooling rate actually observed, i.e., $(\partial T/\partial t)_{rad} >> (\partial T/\partial t)_{obs}$ in many cases. In a practical sense, flux divergence, $\varepsilon = -\partial F/\partial z$, attributable to longwave radiation is usually calculated by the use of radiation charts or special tables or charts (where F is the effective flux of longwave radiation, i.e., the difference between upward and downward fluxes). The charts prepared by Yamamoto and Onishi (1953) are especially suited for this type of calculation because the thermal radiation flux divergence is represented by an area (in polar coordinates). The length of the radius vector is not proportional to dp/dw, the derivative of the transmission function by the contents of the absorbing substance (which is very large for small w and small for large w and would need a very large chart), but to $(dp/dw)^{1/2}$, which has a much smaller range than dp/dw.

By assuming an atmosphere of initially neutral stratification, with a longwave radiation flux divergence that is constant over the period being considered, Gaevskaya et al. (1962) demonstrated (for the case of ΔT_R of -1.78°C/4 hours, where ΔT_R is the temperature change due to longwave radiation flux divergence) the complex factors and mechanisms affecting the cooling of an isothermal atmosphere over 4 hours (see Table 1). ΔT_T and ΔT_{R+T} are, respectively, the temperature change due to turbulent exchange and its combination with the radiative flux divergence.

Table 1 clearly shows that the temperature change ΔT_{R+T} is not simply an addition of changes of ΔT_R and ΔT_T. In the lowest 5 m, the values of ΔT_{R+T} are less than ΔT_T, indicating that, when two kinds of heat transfer mechanisms occur simultaneously, they can have a diminishing effect on each other. The case when ΔT_{R+T} is comparable with ΔT_T or ΔT_R does not prove the dominating influence of either turbulent or radiative heat exchange, because either mechanism of heat transfer, not being a large value, can indirectly influence the other transfer mechanism by its reducing effect. It is understood that the approximation technique presented by Gaevskaya et al. (1962) has not taken into account the interaction of radiative and turbulent heat exchanges and, thus, must be regarded as not logically complete, since the coefficient of mixing is introduced as an external parameter independent of temperature and wind distribution.

In a study primarily interested in the growth process and strength characteristics of surface hoar, Lang et al. (1984) reported temperature profiles very close to the snow surface. Thermocouples were installed at 0.5, 1.0, 1.5, 2.0, 3.0 and 4.0 cm above the snow surface, at the surface (it is hard to determine the actual surface) and at depths of 1.0 and 2.0 cm in the snow. To determine the radiational effects on the temperature reading, they shielded one of the thermocouples with an aluminum cone and reported no significant changes in temperature at any level. The temperature was measured from the evening till early dawn. Since Lang et al. were only interested in the air layer very close to the snow surface (maximum height 4 cm from the snow surface), they reported monotonic temperature increases with height without any indication of a temperature inversion.

In a study related to the visual resolution and optical scintillation over snow, ice and frozen ground, Portman et al. (1961) reported numerous wind speeds and air temperatures taken at 0.5, 1, 2 and 4 m over a 0.5-m deep snow cover. Most of the data were taken at 2-minute intervals from the middle afternoon or early evening to midnight. One set of data covered the early morning hours from 0550 to 0744 c.s.t. and the other set from 0948 to 1030 c.s.t. Since the temperature data (most of them) were reported as the temperature difference between 1 and 0.5 m, 2 and 0.5 m and 4 and 0.5 m, without the temperature at 0.5 m being given, it is impossible to find out the actual temperature at 1, 2 and 4 m. However, from the given temperature difference (ΔT), we still can construct the characteristics of the temperature profile as a function of height. This can be done by noting that, if ΔT is negative, the temperature at 0.5 m is higher. On the other hand, if ΔT is positive, the temperature at 0.5 m is lower. From these limited data, we can conclude that there was a maximum temperature at 2 m.

Table 1. Cooling of an isothermal atmosphere in 4 hours.

Z (m)	ΔT_T (°C)	ΔT_{R+T} (°C)
0	-4.19	-3.76
0.020	-3.11	-2.78
0.048	-2.82	-2.54
0.240	-2.26	-2.06
0.480	-2.05	-1.89
2.000	-1.52	-1.45
4.800	-1.22	-1.20
24.000	-0.71	-0.81
240.000	-0.19	-0.43

In a paper dealing with temperature measurements in an air layer very close to a snow surface, Nyberg (1938) fabricated a temperature measuring device from 0.1-mm-thick platinum wire. He described in great detail the problem of errors caused by resistance and temperature coefficients, as well as the probable errors introduced by radiation. He employed only one resistance thermistor, which could be exactly raised to a specific height (i.e., in increments of 1, 2 and 3 mm, etc.), to measure the temperature profile. Since the air temperature constantly fluctuates, even in the air layer closest to the snow surface, he developed a temperature correction expression to convert the temperature at all heights for the same time on the basis of one measured value.

Nyberg (1938) concluded that the temperature at very stable conditions near the ground could be represented by an exponential law, with the exponent growing linearly with height; he felt that a logarithmic as well as a power law with a constant exponent could not be used to represent the temperature. He also stated that the radiative conductivity, for the air layer close to the ground, was much smaller than the molecular and eddy conductivity. As indicated by Lang et al. (1984), Nyberg (1938) also found no temperature inversions in the air layer close to the snow surface.

Fleagle (1956), with the use of a refractive method, made a theoretical study of the temperature distribution near a cold surface and claimed that his method was superior to conventional methods—these being placing probes at closely spaced intervals—avoiding uncertainties caused by differences in probe fabrication, transfer processes at the probe surface and random temperature fluctuations. His refractive method is based on the vertical deflection of a horizontal light beam passing through horizontally homogeneous air, which can be expressed as follows (Fleagle 1950)

$$h = \frac{x^2(n-1)}{2nT}\left[\frac{g}{R} + \frac{\partial T}{\partial z}\right] + \frac{0.114}{\pi}\frac{\partial \rho}{\partial z}\frac{x^2(n-1)}{2n} \quad (1)$$

where h = apparent height of the source above its true height
x = horizontal distance of the source
n = index of refraction for air
g = gravitational acceleration
R = gas content
ρ = vapor pressure
z = vertical coordinate.

With the assumption that the dew-point and temperature lapse rates have the same order of magnitude, the second term on the right of eq 1 is negligible. The temperature gradient at two heights can be related as

$$\left(\frac{\partial T}{\partial z}\right)_1 - \left(\frac{\partial T}{\partial z}\right)_2 = \frac{2nTz}{xx'(n-1)} \quad (2)$$

from which the difference in temperature gradient can be computed from observing the true horizontal range (x) and the range indicated by the instrument (x'). Subsequently, by repeated observation at a series of heights, the difference in temperature gradients at any two heights may be determined. By numerical integration, Fleagle (1956) computed the temperature profile and reported a temperature anomaly near 10 cm above the cold surface (water). He further indicated that the rate of radiative temperature change—computed from the numerically integrated temperature profile—was rising more than 10°C/hour at the water surface and dropping about 6°C/hour at about 10 cm above the surface. Above 30 cm, the rate of cooling is nearly constant with height and equal to about a 3°C/hour drop. Based on the good agreement between the calculated radiation and the actual observations, Fleagle (1956) concluded that, at a short distance above a cold surface, there exists a small temperature anomaly caused by radiational cooling.

Halberstam and Schieldge (1981) analytically and experimentally studied the anomalous behavior of the atmospheric surface layer over a melting snowpack. They simultaneously measured variations of the net radiative flux, the profile of the wind speed, air temperature and relative humidity above the snow surface, as well as the conductive heat flux and the temperature profile beneath the snow surface. Their experiments were conducted under clear skies, with warm air and calm winds during the day, and cold air and moderate winds at night.

They found that a highly stable sublayer formed near the surface, with a persistent warm layer at about 0.5 m above the snow surface. Though the temperatures at 0.125 m were measured, these data were not shown. They claimed that the air temperature decreased linearly (from its maximum) towards the snow surface. In addition, they also measured humidity at 0.5, 1.0 and 2.0 m. Not only did the humidity increase from early morning to middle afternoon, but it increased sharply between 0.5 and 1.0 m (no data were shown below 0.5 m or between 0.5 and 1.0 m). At or near sunset, these exaggerated gradients of temperature and humidity disappeared, and at night, temperature and humidity had the classical log-linear profiles of stable air.

We would presume that when the surface temperature is significantly lower than that of the air, an intense, stable sublayer would form immediately above the surface, and above that, the air should be fairly isother-

mal (or adiabatic). The reason for the persistent raised maximum at approximately 0.5 m is not clearly understood. De La Casiniere (1974), in a study of heat exchange over a melting snow surface, also reported a thick layer of air apparently reheated by the radiative flux. Granger and Male (1978), in a study of the melting of a prairie snowpack, indicated thermal layering near the snow surface, although not of the magnitude reported by Halberstam and Schieldge (1981). In a study of energy exchange during the melting of a prairie snowpack, Granger (1977) briefly described the warming of the surface snow layer and attributed this phenomenon to radiative flux divergence. The question is, why should such a divergence (or convergence) exist in the lowest layer? Is there any parallel mechanism between this raised maximum to the observed nighttime raised minimum as reported by Lake (1956)?

In Lake's (1956) study, a small unscreened thermistor bead was used along with an ordinary minimum thermometer. Temperature profiles were taken above bare soils under clear nights and stable atmospheric conditions. He found a minimum temperature at heights between 3.8 and 15.2 cm above the surface. The temperature profile exhibited unusual features that were inconsistent with the generally accepted theory of nocturnal cooling of the near-ground air layers. He concluded that the direct radiational heat loss from the air is important in determining the temperature distribution under these environmental conditions. Lake further indicated that both the thermometer and thermistor may have shown temperatures that were different from that of the air owing to radiation exchange, evaporation effects, as well as thermal lag caused by the wind speed. Since the wind speed could not have been greater at the surface than immediately above it, Lake concluded that the air, on a clear night, must lose heat by mechanisms other than convection and conduction to the surface. Therefore, he assumed radiation exchange to be the mechanism for such a loss, and that the air would continue to lose heat until it reached radiative equilibrium with its surroundings.

The shape of the temperature profile Lake (1956) reported is consistent with radiation exchange (Rider and Robinson 1951), but the surface temperature was unusually high compared with that of the air. The suggestion of an influx of cold air from the surrounding fields during the early part of the night cannot account for the continued cooling of the air during the calmest part of the night without the contribution of radiative heat loss from the air. The density inversion in the lowest layers might be expected to break down rather easily and thus may provide the explanation of the fluctuations in inversion conditions observed by a number of researchers. The steady conditions of moderate or large inversions are rare, and the normal behavior is more or less a periodic buildup followed by a rapid breakdown with a period on the order of 0.5 to 1 hour. However, Lake's data, taken under exceptionally calm conditions, showed no evidence of such a periodic nature. Another important fact that is related to this phenomenon is that the intensity of temperature fluctuations are known to be closely related to the temperature gradient. The temperature fluctuations become less and tend to disappear as the gradient becomes small. Therefore, the standard deviation of temperature about the mean had a relatively higher value close to the surface than that at the level above the surface having the minimum temperature.

The importance of longwave radiative flux in air layers near the ground has been long recognized; it has been a common practice to estimate the radiative flux divergence either from radiation charts or by direct methods, using temperature and humidity in the lowest 100 m. However, it has been a common belief that direct measurement was difficult, as net radiometers were neither sensitive nor stable enough to measure small differences with sufficient accuracy.

Funk (1960), using an improved polyethene-shielded net radiometer, directly measured nocturnal radiative flux divergence profiles in the lowest few meters. He reported that, between 0.5 and 1.5 m for a wide range of wind speeds and cloud conditions, the radiative flux calculated from radiation charts is lower and up to only one-third of his measured values. He attributed the discrepancy to haze. Since the radiative cooling rates in the lowest few meters were found to be higher than actual rates, he believed that the development of the nocturnal inversion near the ground was largely driven by radiation, with turbulent transfer often acting as a brake.

Geiger (1965) cited this phenomenon, which has been reported by many investigators. Oke (1970) verified this and suggested that longwave radiative flux divergence and a discontinuous humidity profile caused heat loss from an elevated layer. Zdunkowski (1966), however, indicated that longwave radiative emission from a layer of water vapor could not account for the raised maximum reported by Halberstam and Schieldge (1981). Instead, he proposed that a haze or fog layer could act as a stronger radiator and could, thus, cool a level above the surface. But this hypothesis does not fit into Halberstam and Schieldge's work because they did not observe any significant fog or haze and the raised maximum persisted well into the afternoon when the fog or haze would have evaporated, even if they had been present earlier in the day. They, however, noted that there was an increase in humidity towards the snow surface, suggesting that there may have been an interac-

tion between the humidity and temperature.

To verify that their measured humidity profile could create sufficient longwave radiative flux divergence for the raised maximum, Halberstam and Schieldge used the measured humidities' values at 0.5, 1.0 and 2.0 m and the formula given in Kondrat'yev (1969) to calculate the steady-state temperature at these levels.

$$\frac{\partial T}{\partial t} = \frac{q}{c_p}$$

$$\left[\int_{p_0}^{p} \frac{\partial \tau}{\partial w}\frac{\partial B}{\partial p}\,dp - B_\infty\left(\frac{\partial \tau}{\partial w}\right)_\infty - \int_{p}^{p_T} \frac{\partial \tau}{\partial w}\frac{\partial B}{\partial p}\,dp\right] \quad (3)$$

where c_p= heat capacity of air
B = Planck function
τ = transmission function
w = precipitable water vapor
p = atmospheric pressure at the level in question
p_0 = atmospheric pressure at $z = 0$
p_T = arbitrary pressure at the top of the atmosphere.

The function τ is given as the sum of four wavelength bands defined as

$$\tau = \frac{1}{4}\sum_{i-1}^{4} e^{-k_i w} \quad (4)$$

where k_i is the absorption coefficient for water vapor in distinct wavelength bands. Using the relations

$$\frac{T}{T_0} = \left(\frac{p}{p_0}\right)^{\gamma} \text{ and } \frac{q}{q_0} = \left(\frac{p}{p_0}\right)^{a}$$

and two cases of $a = 2.0$, $\gamma = 0.286$ (adiabatic) and $a = 1.2$ and $\gamma = 0.2$, Halberstam and Schieldge numerically integrated eq 3 and reported that, in both cases, equilibrium ($\partial T/\partial t < 10^{-7}$°C) was reached at near-isothermal conditions. They concluded that the humidity profile alone will not produce the pronounced local maximum.

The other important atmospheric heat is the solar radiation (especially in the absence of strong wind). In general, shortwave radiative flux divergence is considered to be negligible in the lower atmosphere, but Faraponova et al. (1968) claimed that this is not the case. At Halberstam and Schieldge's experimental site, the atmosphere is humid near the snow but quite dry aloft, and the elevation of the site caused intensive solar radiation. Furthermore, the high albedo of snow nearly doubles solar radiation in the air near the surface.

To include both the effect of solar radiation, as well as the longwave radiative flux, on the temperatures distribution, they estimated the solar flux at any level by

$$F_s = S_0(1 - \beta)\cos\hat{Z}$$

where S_0= solar constant
$\hat{Z}$ = solar zenith angle (a function of solar declination δ, latitude ϕ and hour angle H)
β = absorptivity of the column directly above the level in question.

With the use of β given by Washington and Williamson (1977) as

$$\beta \cong 0.11\left(w\sec\hat{Z} + 6.31\times 10^{-4}\right)^{0.3} - 0.0121 \quad (6)$$

and with the site close to the equinox (therefore, $\delta = 0$) $\cos\hat{Z}$ can be written as

$$\cos\hat{Z} = (\cos\phi)(\cos H) \quad (7)$$

Using a value of $S_0 = 900$ W/m^2, they calculated the temperature at 0.5, 1 and 2 m and found that a raised maximum existed during daylight at 1 m and that there was a large difference between the snow surface and the air. Halberstam and Schieldge concluded that the warm layer at about 0.5 m over the snowpack was a result of the combined effects of light winds, strong solar radiation and the presence of a snow surface. They proposed the following mechanism as the essential process in the formulation of the warm air layer during the sunny periods: snow sublimes into the atmosphere and, because of the high stability and low wind velocity, the water vapor remains concentrated in the lower surface layer. In addition, because of the dry upper atmosphere and the elevation of the site, the solar radiation is mostly absorbed in the lowest levels in both downward and upward directions. Furthermore, the layers of water vapor nearest the surface lose their heat because of longwave radiative emission and weak turbulent mixing; on the other hand, the upper levels also lose more heat than the middle layers owing to less absorption of solar radiation and more longwave radiation to space. The appearance of the raised maximum ends as soon as the sun sets, because the atmosphere no longer receives heat from the sun, and the snow surface cools rapidly and ceases sublimation. Subsequently, a stable layer is formed that conforms to the classical log-linear form.

The phenomenon of the raised maximum may play an important role in the energy exchange between the snow surface and the atmosphere. Instead of creating a stable atmosphere and retarding the turbulent motion, snow can transform the incoming solar radiation into a source of heat, not by convective motion, but through supplying moisture and reflected sunlight to the atmosphere. The heated atmosphere would then act as an elevated heat source and, like a blacktop surface in the

presence of strong radiation, cause unstable profiles and upward convection. Therefore, detailed knowledge of the temperature profile near the surface would be required to correctly estimate the sign and magnitude of turbulent transfer. The warmed atmosphere would also radiate in the longwave to sustain the snowmelt, despite the presence of a high albedo (0.95 for fresh snow), weak winds and a high emissivity (~0.99). A model developed by Halbestram and Melendez (1979) indicates that this heat transfer mechanism maintains the snow surface temperature at 0°C instead of allowing it to decrease.

Granger and Male (1978), however, reported that heat transfer to the snow is highly associated with the air temperature at 85.0-kPa atmospheric pressure. This implies that the air mass temperature is more important in convective transfer than the temperature profile right above the snow surface.

A number of investigators (e.g., Anderson 1976, Jordan 1991) have worked on the energy balance at the snow/air interface. Studying the energy balance requires the determination or prediction of the turbulent fluxes of sensible and latent heat, the net radiation flux, heat flux from precipitation and the geothermal flux from the ground. Net radiation flux can be measured by radiometers. The turbulent fluxes are usually estimated from measurement of profiles of air and dew-point temperatures and wind speeds at reference levels, with the assumption that vertical flux is constant with height. Panofsky (1974) extended the constant-flux layer up to 30 m, but indicated that the height can be quite variable and is influenced by strength of the wind, the temperature gradient and upwind surface conditions.

Although the concept of the constant-flux layer is a convenient one for measuring or calculating the turbulent fluxes, Busch (1973) pointed out that the thermodynamic energy equation for boundary layer flow includes a term representing the rate of temperature change ascribable to the divergence of radiant heat transfer. Therefore, the usual assumption of insignificant contribution from this flux divergence in micrometeorological applications (a few meters above the surface) is questionable and is a major source of error in experimental results. Male and Granger (1981) indicated that, in the case of melting snow, and if the air mass is cool, the raised temperature maximum (as reported by Halberstam and Schieldge 1981) in the air layer 0.2 to 0.5 m above the snow surface will split the heat flow in opposite directions. That is, above the raised maximum, the temperature profile is unstable, causing heat flow upward, away from the surface; below the raised maximum, heat flow is down toward the surface. Therefore, the heat flux is not only not a constant with height but reverses.

Brunt (1929) analytically studied the problem of heat transfer by radiation and turbulence in the lower atmosphere. With the assumption that a column of air, containing 0.3 mm of precipitable water in the form of vapor, will completely absorb all the radiation of wavelength between 5.5 and 7 μm and above 14 μm, the net upward flux of radiation can be written as

$$q_{r,n} = -k \frac{\partial T}{\partial z} \tag{8}$$

where k is defined as a product of

$$\frac{\partial E}{\partial T} \text{ and } \frac{bT}{p_w}$$

($b = 0.03\ R_w$, where R_w is the gas constant for water vapor) in which E is the radiative energy, T is the absolute temperature, and p_w is water vapor pressure in millibars. In general, the quantity k varies slightly with height. The vapor pressure p_w varies from day to day at a given place but usually shows no marked diurnal variation even though it varies to some extent with height. The value

$$\frac{\partial E}{\partial T}$$

can be expressed as

$$\frac{\partial E}{\partial T} \times 10^3 = 3.0 + 0.2\,(T - 270) \tag{9}$$

and for $T = 275$ K and $p_w = 5$ mbar, the value of k is approximately 2.18×10^2 W/m K. Assuming the lapse rate of temperature to be adiabatic (i.e., $\partial T/\partial z = -10^{-2}$ [°C/m]), we see that the vertical flow of radiation will be

$$q_{r,n} = -k \frac{\partial T}{\partial z} = 2.18\left(W/m^2\right) .$$

Consider a disk of air with unit horizontal area, and treat the problem like the one-dimensional heat conduction in solids. The heat gain by air can be written as

$$\frac{\partial T}{\partial t} = K_R \frac{\partial^2 T}{\partial z^2} \tag{10}$$

where

$$K_R = \frac{bT}{\rho c_p\, p_w} \cdot \frac{\partial E}{\partial T}$$

and for $T = 275$ K and $p_w = 5$ mbar, the value of K_R is $\sim 1.7 \times 10^3$ (cm^2/s), which is considerably higher than the thermal diffusivity (i.e., $\alpha = 0.16$ in the same units). For the case of a layer less than ℓ in depth near the ground, the expression for $q_{r,n}$ takes the form of

$$q_{r,n} = \frac{k}{2} \frac{\partial T}{\partial z} \tag{11}$$

and the net upward flux is

$$q_{r,n} = -\frac{\partial E}{\partial T}\frac{\partial T}{\partial z}\frac{\ell^2 + 2\ell z - z^2}{2\ell} \qquad (12)$$

which reduces to eq 11 when $z = 0$ and to eq 8 when $z = \ell$. Thus, the value of

$$k\left(=\ell\frac{\partial E}{\partial T}\right)$$

decreases steadily from $z = \ell$ to half that value at the ground. Therefore, for conditions of steady state, so that the upward flux remains the same for all heights, the lapse rate $\partial T/\partial Z$ at height ℓ must be half of its value at the ground. This required condition of steadiness is possibly reached in inversions on clear nights.

Dealing with the transfer of heat by turbulence, Brunt (1929) derived the resultant upward eddy flow of heat across the isobar p as

$$q_{tp} = -K_E \rho c_p \left(\frac{\partial T}{\partial z} + \alpha_{da}\right) \qquad (13)$$

where K_E is defined as

$$K_E = \frac{\Sigma m (p_0 - p)}{g\rho^2}$$

and p_0 = pressure when the eddy breaks away carrying the original potential temperature

$\alpha_{da} = \frac{g\gamma}{R}$; the dry adiabatic lapse rate

γ = ratio of the specific heat at constant pressure to that at constant volume.

The net flow of heat is upward, downward or zero, depending on whether

$$\frac{\partial T}{\partial z}$$

is greater than, less than or equal to α_{da}. The net heat gain between the two isobar is

$$\rho\frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left(K_E \rho\left(\frac{\partial T}{\partial z}\right) + \alpha_{da}\right) \quad . \qquad (14)$$

With the further assumption of constant ρ and K_E, eq 14 becomes

$$\frac{\partial T}{\partial t} = K_E\frac{\partial^2 T}{\partial z^2} \qquad (15)$$

which is directly comparable with that of conduction of heat in a solid. Brunt (1929) went a step further and reported that a reasonably good first approximation to the temperature variation produced by radiation and turbulence can be expressed by

$$\frac{\partial T}{\partial t} = (K_E + K_R)\frac{\partial^2 T}{\partial z^2} \quad . \qquad (16)$$

Both K_E and K_R are positive. But, he stated that it is very probable that K_E and K_R will be found to vary differently with height and time of the day. Vapor pressure is considered a more important factor in causing the variation of K_R. Since the vapor pressure does not have a pronounced diurnal variation, K_R will vary within only a limited range with time of day. On the other hand, K_E is likely to vary within a wide range according to the changes of the lapse rate, being smaller under stable rather than unstable conditions. By treating both K_E and K_R as constant values over a small range of height, the rise or fall of the temperature depends on the positive or negative values of $\partial^2 T/\partial z^2$. The temperature change with height attributable to radiating heat transfer alone is determined by the direction of radiation (it decreases with upward radiation and increases with downward radiation). On the other hand, the direction of heat transfer by turbulence is determined by the lapse rate. If the lapse rate is greater than the adiabatic lapse rate, the heat flow is upward; it is downward if the lapse rate is less than the adiabatic lapse rate.

The values of K_E vary widely and have been estimated by numerous researchers from observations of temperature or wind at different heights. K_E has a value ranging from 3×10^3 to over 10^5 cm^2/s. The K_R value is comparable with that of K_E in the case of an inversion, but is much smaller under conditions of vigorous turbulence. In fact, eq 16 can be used to derive the sum of K_E and K_R if observations of temperature at different heights are available. If the sum is considerably in excess of 10^3, we can neglect the effect of K_R and consider the sum as being the value of K_E alone.

In a subsequent paper, Brunt (1930) pointed out the constancy of the mean lapse rate of temperature at all heights within the troposphere and in all latitudes. The variation about the mean value (i.e., one-half of the dry adiabatic lapse rate) is very slight at all heights greater than a few hundred meters above the ground, but is greatest in the layer nearest to the ground. At night and especially during clear nights in winter, there is a sign change of the lapse rate in the lowest layer, i.e., the temperature increases with height instead of decreasing. On sunny summer afternoons, the lapse rate in the lowest layers attains very high values, i.e., the change of temperature from 0.5 to 1 m above the ground can be 100 to 200 times the dry adiabatic lapse rate. Brunt (1930), with the assumption of black body radiation at 280 K, which gives 2.0×10^2 W/m^2, and with the use of eq 11, demonstrated that the value of dT/dz can be calculated from

$$\frac{1}{2}k\frac{\partial T}{\partial z} = 2.0 \times 10^2 \left(\mathrm{W/m^2}\right)$$

where k is $1.36 \times 10^2/p_w$ (W/m K). The value of $\partial T/\partial z$ is approximately 0.0030 p_w or approximately 30 p_w times the dry adiabatic lapse rate (i.e., $-10^{-4}\partial T/\partial z$ °C/cm). If $p_w = 10$ mbar, then the maximum $\partial T/\partial z$ is on the order of 300 times -10^{-4}. In the case of high sun in a clear sky, the incoming radiation may attain more than double the value used in this calculation. Therefore, the maximum $\partial T/\partial z$ will be on the order of 600 times -10^{-4} (°C/cm) or close to about 0.06°C/cm. In the above analysis, it should be pointed out that I have not taken into account the effects of turbulence as well as the layer stability and have not spelled out whether it is dynamically possible to establish such a large lapse rate.

With the assumption of invariant K_E and K_R as well as, initially, that the lapse rate is constant, i.e., $T = T_s + \beta z$ with height, Brunt (1930) solved eq 16 for two broad cases. In the first case, the surface temperature changes uniformly at ω degrees centigrade per second or $T_s = T_0 + \omega t$; in the second case, the temperature at the surface changes from T_s to T_s' at $t = 0$ and remains at T_s' afterwards. With the introduction of the dimensionless variable

$$\zeta^2 = \frac{z^2}{4(K_R + K_E)t}$$

the solution for the first case is

$$T = T_s + \beta z + \omega t\left[\left(1 + 2\zeta^2\right)\left(1 - \mathrm{erf}\zeta\right) - 2\pi^{-1/2}\,\zeta e^{-z}\right] \tag{17a}$$

or

$$T = T_s + \beta Z + \omega t \Psi(\zeta) \quad . \tag{17b}$$

For the second case, the solution is

$$T = T_s + \beta z + \left(T_s' - T_s\right)\left(1 - \mathrm{erf}\zeta\right) \tag{18a}$$

or

$$T = T_s + \beta z + \left(T_s' - T_s\right)\phi(\zeta) \tag{18b}$$

For both cases, when $\zeta = 1$, the values of $\Psi(\zeta)$ and $\phi(\zeta)$ are, respectively, on the order of 0.06 and 0.16, and, therefore, the effect of the variation of the surface temperature is negligible beyond the height corresponding to $\zeta = 1$ or $z^2 = 4(K_E + K_R)t$. For the case when only radiation is effective (i.e., under very stable conditions with no turbulence, then $K_E \rightarrow 0$) and for an average value of $K_R = 650$ cm^2/s, then $z = \sim 50\, t^{1/2}$ cm. We can see from this expression that the effects of radiation are limited to a shallow layer near the surface (up to 100 m in 11 hours). Therefore, the major diurnal change of temperature in the upper air cannot be regarded as caused by radiation transmitted from the ground upwards. On the other hand, for the case of $K_E + K_R = 10^5$ cm^2/s (or $K_E \approx 10^5$, since the maximum value of $K_R \equiv 10^3$) corresponding to a fairly turbulent condition, it takes only about 250 seconds to transmit the change of surface temperature to a height of 100 m.

The temperature change caused by a sudden appearance of the sun from the clouds can be predicted from eq 18b. In this case, the ground and the air in contact with it will increase to a higher temperature almost instantaneously, and will remain fairly constant afterwards. The expression $z^2 = 4(K_E + K_R)t$ can be used for determining the lag in the rising of temperature at different heights, as well as for evaluating the sum of $K_E + K_R$ if the times t_1 and t_2 for commencement of the rise of temperature at heights z_1 and z_2 are available.

The results computed from eq 18b were verified by solving eq 16 with the boundary condition of $T = A \sin(2\pi t'/24)$. The solution is

$$T = Ae^{-bz}\sin\left(\frac{2\pi t'}{24} - bz\right) \tag{19}$$

where A is the amplitude and b has a value of

$$\left[\frac{\pi}{\left[24(K_E + K_R)3600\right]}\right]^{1/2}.$$

In case of $K_E \rightarrow 0$ and $K_R = 650$, b has a value of 2.4×10^{-4} cm^{-1}. The amplitude of the diurnal variation of temperature will be reduced to 1/100 of the surface value at a height of 200 m, and the time of maximum temperature at this height will be about 18 hours after the time of maximum temperature at the surface. The effects of radiation alone are only of importance in the layer in immediate contact with the surface of the Earth, and the large-scale variation of temperature with height cannot be explained as purely radiational effects. In deriving these results, Brunt (1930) assumed that all radiation is either completely absorbed or completely unabsorbed in any layer containing 0.3 mm of precipitable water.

The formation of nocturnal inversions was also discussed by Brunt (1930). Equation 17b can be used by changing the sign of the uniform change rate ω to calculate the propagation of the cooled air layer upward. As indicated, in the case of the expansion of the heated-air layer by radiation and turbulence, the inversion of the lapse rate at the ground extends 30 m in 1 hour to about 100 m in 11 hours (see Table 2). He also concluded that these values are in good agreement with observations.

As indicated in the above, not many detailed experiments have been conducted solely to measure the temperature profile over a snow surface to verify the existence of a temperature inversion in the air layer near that surface. The work of Halberstam and Schieldge (1981) is the most comprehensive and the most up to date

Table 2. Upward expansion of the heated-air layer (after Brunt 1930).

Depth of heated layer	$K_R + K_E = 10^5$	$K_R + K_E = 10^3$	$K_R + K_E = 650$
50 cm	0.006 s	0.6 s	1 s
1 m	0.025 s	2.6 s	4 s
5 m	0.625 s	1 min	100 s
10 m	2.5 s	4 min	6.66 min
30 m	22.5 s	37 min	1 hour
100 m	250 s	400 min	11 hours

investigation over melting snow. They claimed that there is a distinct temperature rise from 0°C at the snow surface up to a maximum temperature at about 0.5 m. The temperature then decreases as the distance from the snow surface increases.

Lang et al. (1984) conducted similar experiments, but carried out their work in the evening and early morning, and their interest was limited to a few centimeters above and below the snow surface. They found no abnormal temperature distribution within the region of interest.

De La Casiniere (1974) also measured temperature as well as wind velocity over a melting snow surface at 0.25, 0.5, 1 and 2 m. The data from this study show a continual temperature increase from the surface to a maximum at about 0.25 m from 0900 to 1500. The temperature then decreases as height increases up to approximately 0.7–0.8 m and then reverses the trend, increasing gradually up to 2.0 m. However, in this study, there was no measurement of the temperature between the surface and 0.25 m. So, there is no way to tell whether there is any temperature reversal within this layer, and the temperatures of the snow surface and the air layer in contact with it are vital to verifying the commonly accepted constant-flux layer extending from the surface to up to about 30 m.

In most of the current work on the energy balance over snow (i.e., Anderson 1976, Jordan 1991), the sensible heat flux is estimated based on the concept of a constant flux with height. As we all know, the atmosphere is in constant motion, and its temperature is connected directly or indirectly to the energy sources of solar radiation, longwave radiation and air turbulence. It is highly probable that the air temperature nearest to the snow surface layer will exhibit special features caused by its limiting temperature at 0°C and its continuous supply of water vapor, which absorbs solar radiation in both directions (i.e., incident and reflected). Subsequently, warming of the air near the surface results in the overall formation of the raised maximum and the phenomenon of a temperature inversion.

The purpose of this study was to measure the temperature profile starting from or near the snow surface and up to about 1.3 m. The field measurements were made in the presence or absence of snow- and ice-covered surfaces. Simultaneous sonic measurements were made to derive any meaningful correlation between the temperature profile data and the sensible heat flux data. I expect that if there are special features in the temperature profile determined during the winter in the presence of a snow cover, the current widely used method of predicting the sensible heat flux for the energy balance will have to be modified.

EXPERIMENTAL

Site selection

We spent a great amount of time attempting to secure a site that was logistically acceptable and economical, as well as suitable for micrometeorological measurements. For sonic measurements, the horizontal homogeneity of the area upwind of our instruments should be uniform for a fetch that is roughly 100 times the measurement height if we are to avoid nonrepresentative measurements introduced by internal boundary layers. However, even after considerable effort by all the research investigators conducting experiments on a snow cover, we could not find a feasible site. Finally, we decided on a site at the back of the main CRREL building. The site is a rather small rectangle (about 8 × 8 m) that is elevated on three sides and bordered on one side by a similar, but smaller, rectangular cavity. All around there are all sorts of obstacles, such as trees, buildings, a hut and plants, that offer resistance to the air flow and produce extra air turbulence that contrasts with a normally homogeneous field having a large and smooth upwind fetch. In the beginning, we thought that this site, at least, would provide a test for the workability of our sonic instruments employed for sensible heat measurements. As times goes by, we simply do not have many alternatives as far as the experimental site is concerned. In addition, I wanted to conduct the sonic and temperature profile measurements concurrently, so I decided to have the temperature profile device installed near the sonic measurement tower.

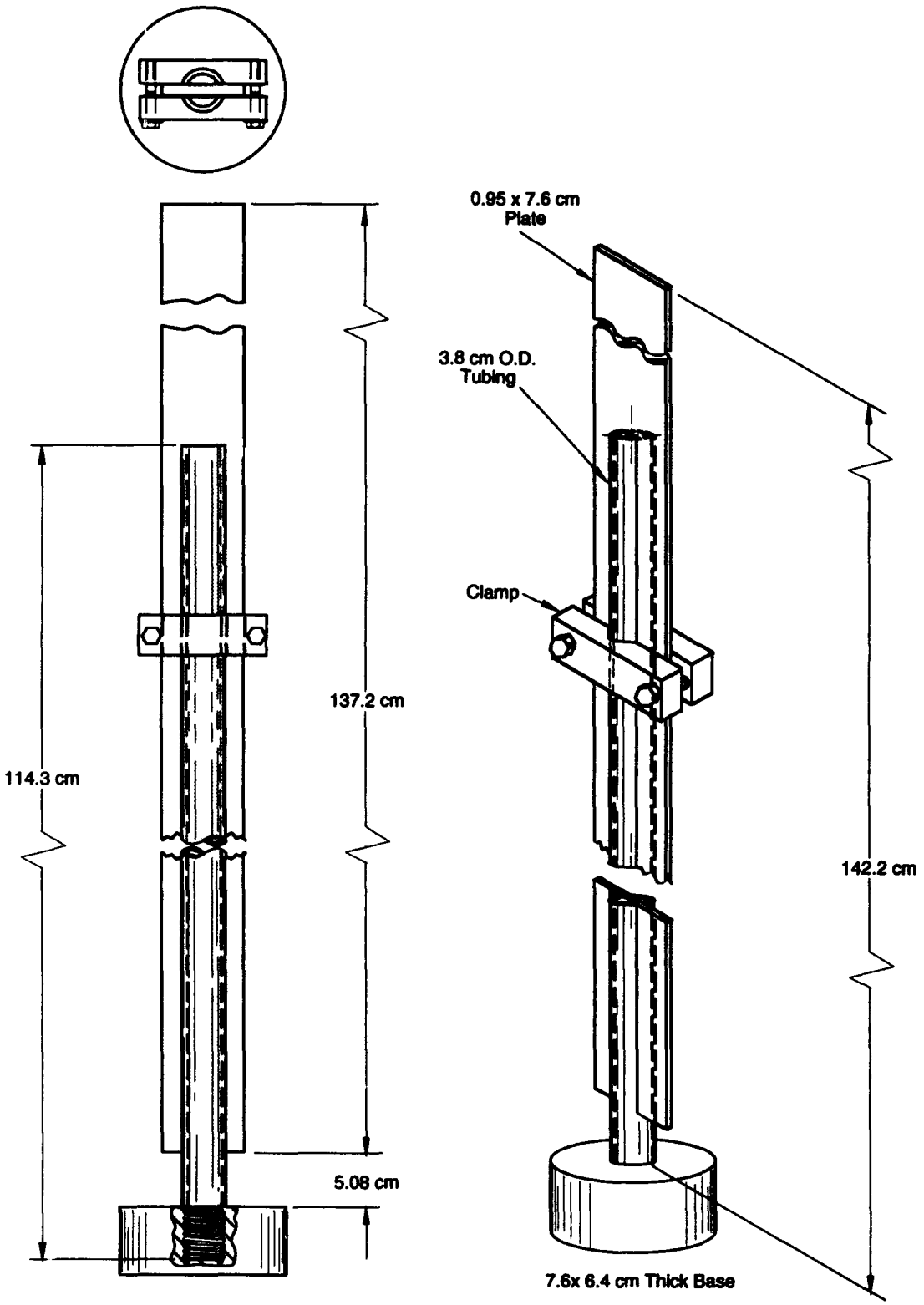


Figure 1. Temperature profile measuring device.

Experimental measurement

A simple device was designed and built to install eight copper-constantan 20-gauge thermocouples spaced at 1, 3, 5, 9, 17, 33, 65 and 129 cm from the bottom of a rectangular Lucite bar that can slide up and down. Figure 1 shows the schematic of the setup. Since my interest is in temperature measurements above the surface, after a snow storm I can slide the Lucite bar upward without altering the spacing of the thermocouples, resetting the first thermocouple to about 1 cm above the snow surface. The entire device can also be rotated to a certain extent to adjust to the prevailing wind direction.

All the thermocouples used in this experiment were calibrated in an ice bath and proved to be accurate to within ± 0.1°C.

A Campbell CR7 data logger was programmed to take the eight thermocouple readings at sampling rates of 30 seconds, 1 minute or 2 minutes, with the selection of data averaging time ranging from 5, 10, 20, 30 to 60 minutes. The temperature readings were expected to vary with the time of day, amount of sunshine, wind velocity and season of the year. When weather permitted (in case of strong wind, rain or snow, the sonic instrument cannot be used), both the sonic and temperature profile measurement were always made concurrently.

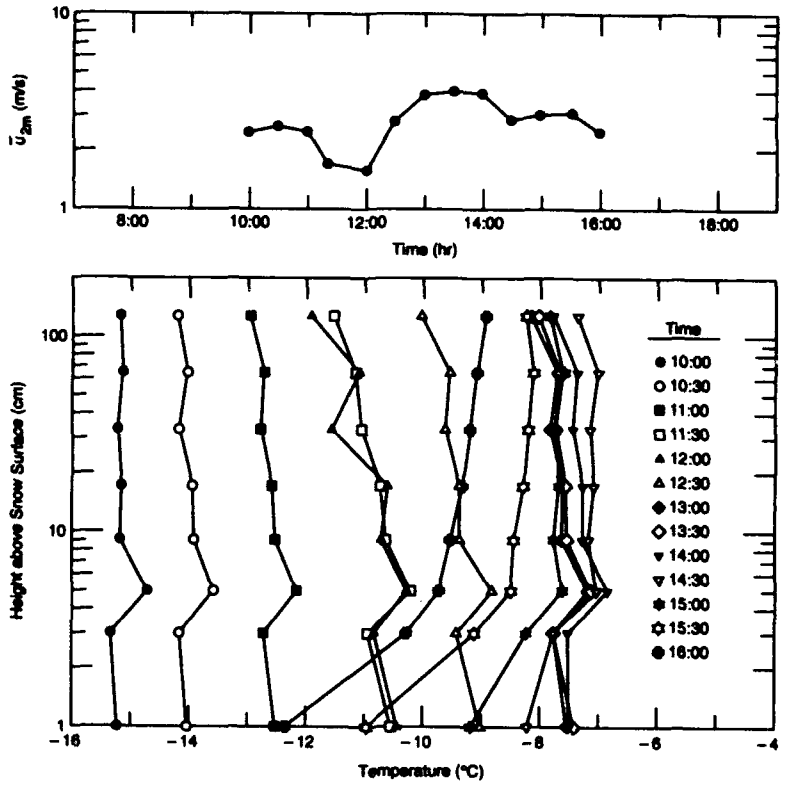
RESULTS AND DISCUSSION

Between 20 December 1991 and 7 May 1992, 24 temperature data sets were obtained. The data recording begin usually in the morning at 0900 to 1000 e.s.t. and ended at 1500 to 1700. In most cases, the temperature data are averaged over 10 minutes. The data were taken under a wide variety of meteorological conditions (i.e., completely and partially sunny, high clouds, etc.), but no data were taken during snow, rain or on very windy days. These preliminary data are presented in semi-log plots to show their variations with time and height above the ground (Fig. 2). On each day the shape of the temperature profile did not vary much in terms of height or time of the day as long as these data were taken during sunshine. In fact, in most cases (regardless of the presence or absence of the raised maximum temperature), the semi-log plots are more or less linear with height and the temperature decreases slightly as height increases.

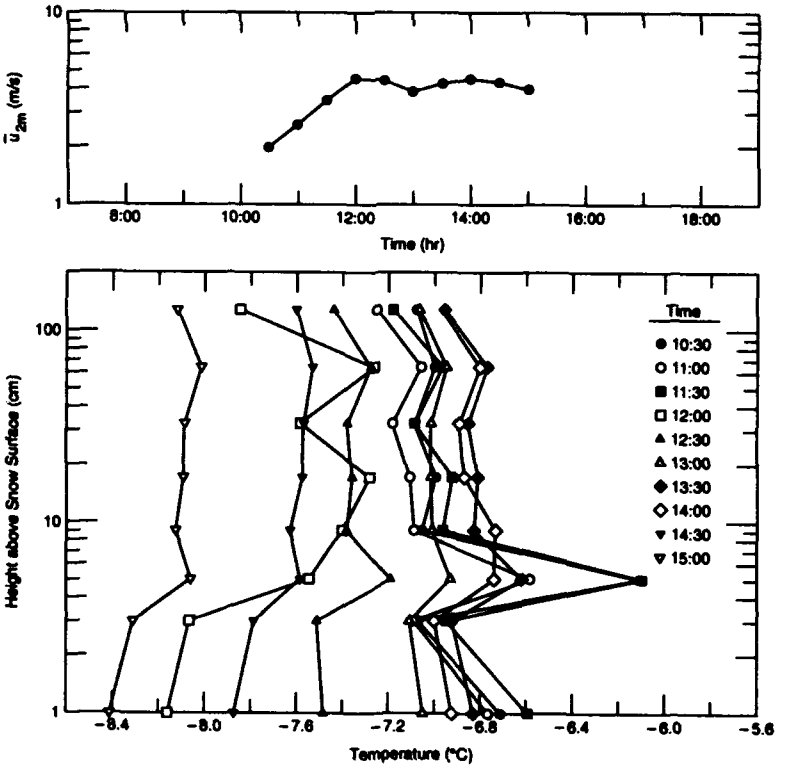
Figure 2a shows the temperature distribution over a fluffy snow cover that was only 1 day old, of very low density and about 10 cm deep. It was mostly sunny, with moderate wind speeds, and occasional clouds passed over between 1100 and 1200. The temperature of the air (above the snow surface) had been increasing almost at the same rate from the snow surface up to 1.28 m from 1000 to 1100, but from 1130 to 1200, except at heights of 0.33 and 1.28 m, the two profiles more or less overlap. During this period the wind speed at 2 m $\overline{U}_{2m}$ was nearly constant, so the convective turbulent effect on the temperature reading should have been similar. However, the temperature profile at 1230 resumes the shape it had at 1100, but the air temperatures throughout the height increased from −14 to −10°C. We can also see that the largest $\overline{U}_{2m}$ increase was during the period from 1200 to 1300, accompanied by the largest increase in temperature that maintained profiles similar to those at 1000. After 1300 the temperature changed very slowly because the net solar radiation intensity decreased from about 150 to 10 W/m^2 by 1430. (Many factors affect the temperature readings: the direct and reflected solar radiation in visible and near IR [0.3–3 μm] is one of the major factors during the day; wavelengths in the 3–50 μm range will play a major role during the night. The sole purpose of using solar intensity values is to find any corresponding temperature reading change.) Soon after sunset (about 1600), the familiar form of a stable air layer appeared, i.e., colder air at the surface and warmer at the top, with a great portion of the air layer (that covered by the temperature measurement setup) increasing its temperature very gradually.

The most remarkable feature of the temperature profiles, as the time progressed from the start of the experiment to sunset, is the temperature reversal near the snow surface, i.e., there is a slight decrease in temperature from the surface to 2 cm and then an increase from 2 to 5 cm. After reaching the maximum at 5 cm, the temperature reverses again, decreasing gradually as height increases. As far as I know, no one has ever reported this double-reversal of temperature over a snow surface (for snow temperature either at the melting point or below it). To ensure the reality of this temperature-reversal phenomenon, thermocouples at the 8- and 2-cm levels were switched with those at the 4- and 1-cm levels respectively. There was no change in the observed temperature profile. I believe that this temperature reversal is not caused by the temperature measuring device but by the complex microphysical heat exchange processes occurring near the snow or snow- and ice-covered surfaces.

The temperature profiles were not recorded overnight for most of the cases because the thermocouples would have been covered with a cylindrical shell of frost as evening progressed, especially during clear and cold nights. Therefore, they would not have accurately measured air temperature but the temperature of the

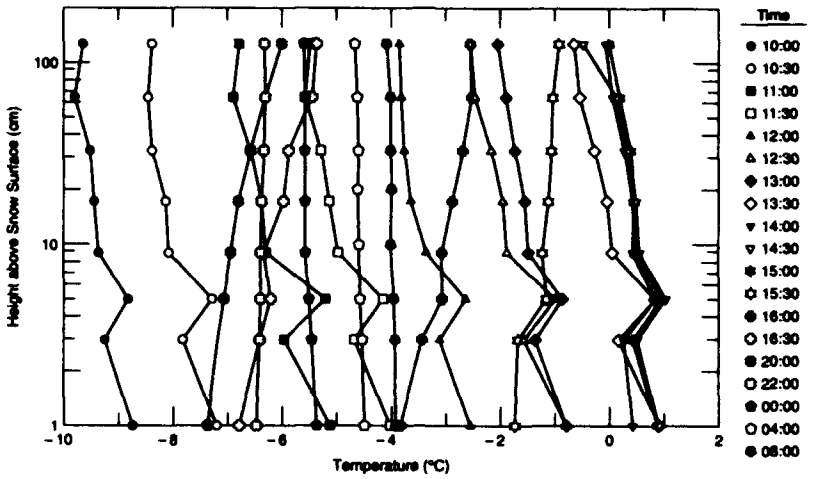


a. Over a fluffy snow surface.

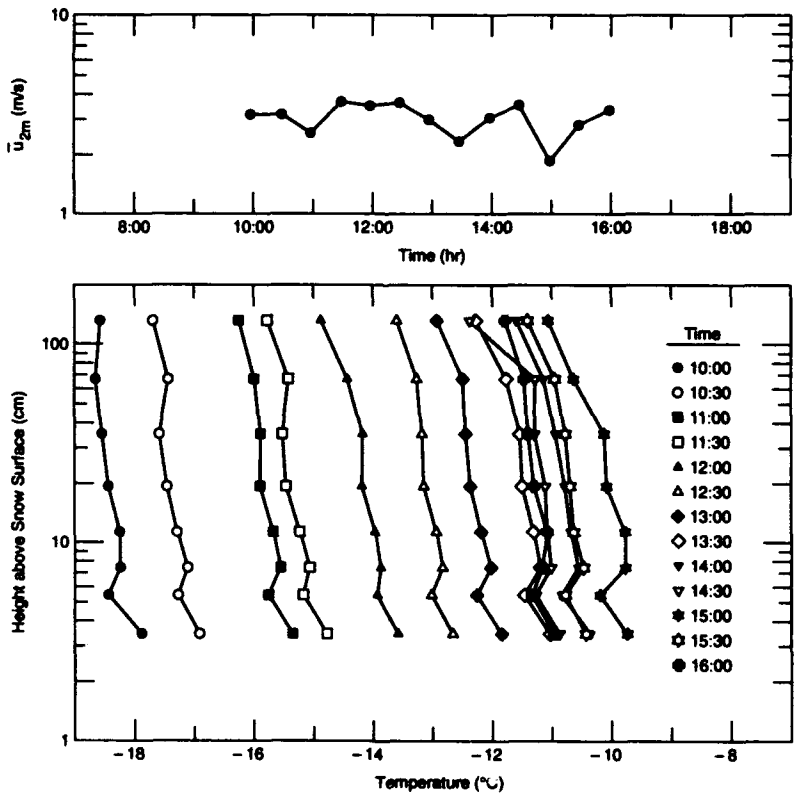


b. Over a snow surface.

Figure 2. Temperature and $\overline{U}_{2m}$ variations with height as function of time.

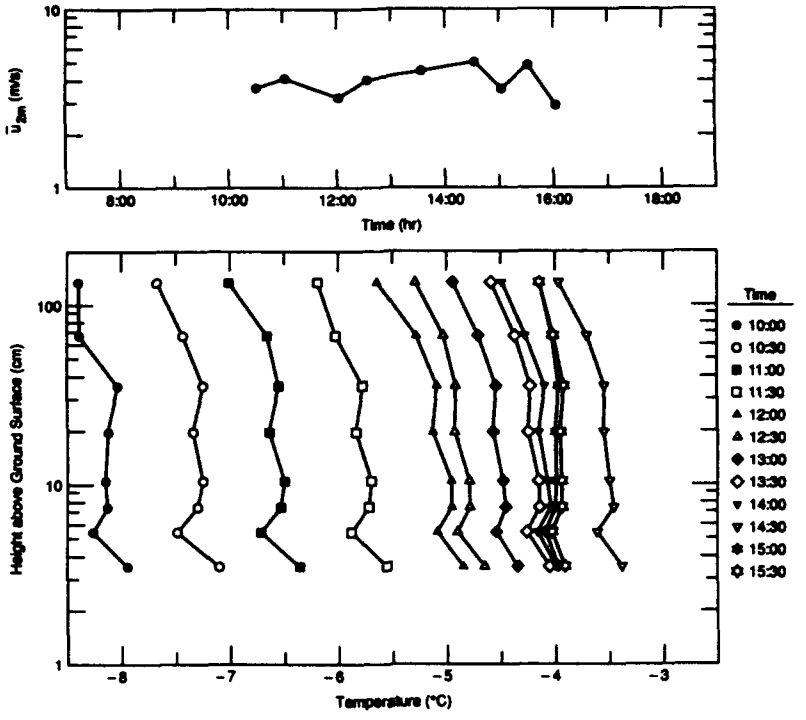


c. Over a hard-crusted snow surface.

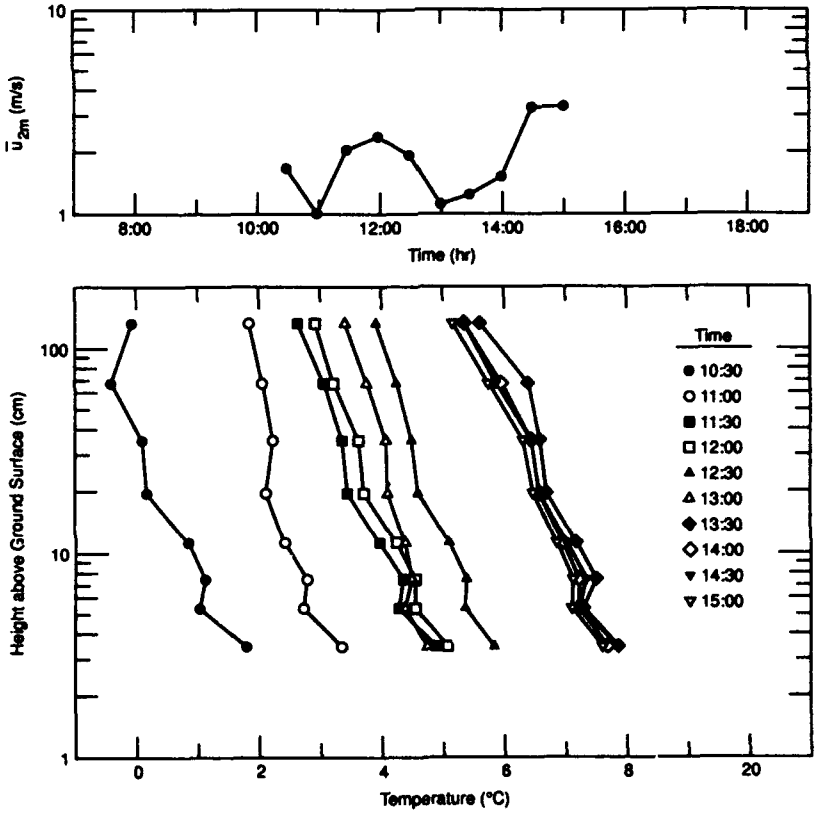


d. Over a thin layer of snow and ice.

Figure 2 (cont'd).

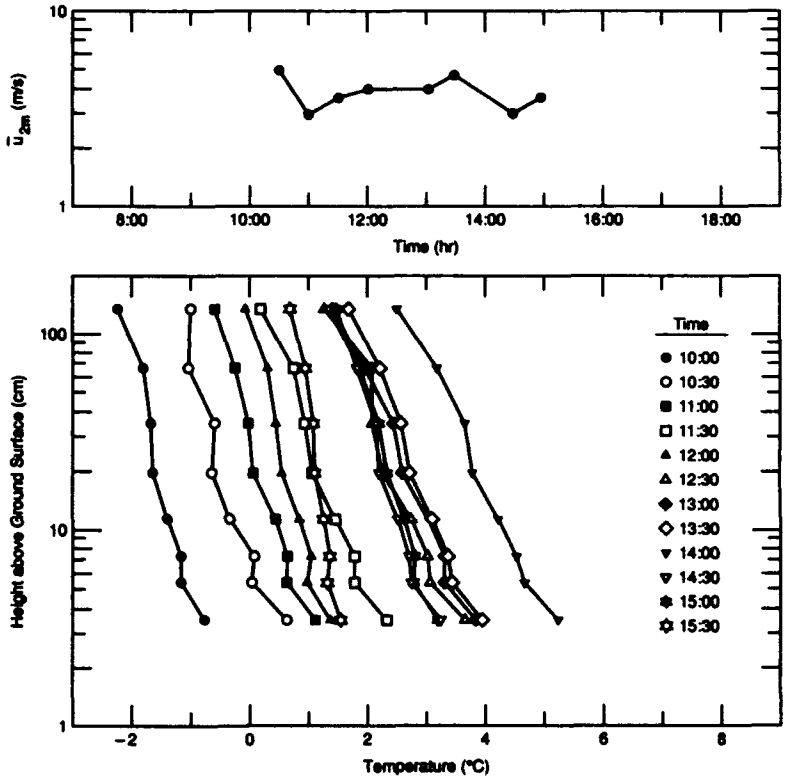


e. Over a discontinuous snow (or ice) cover.

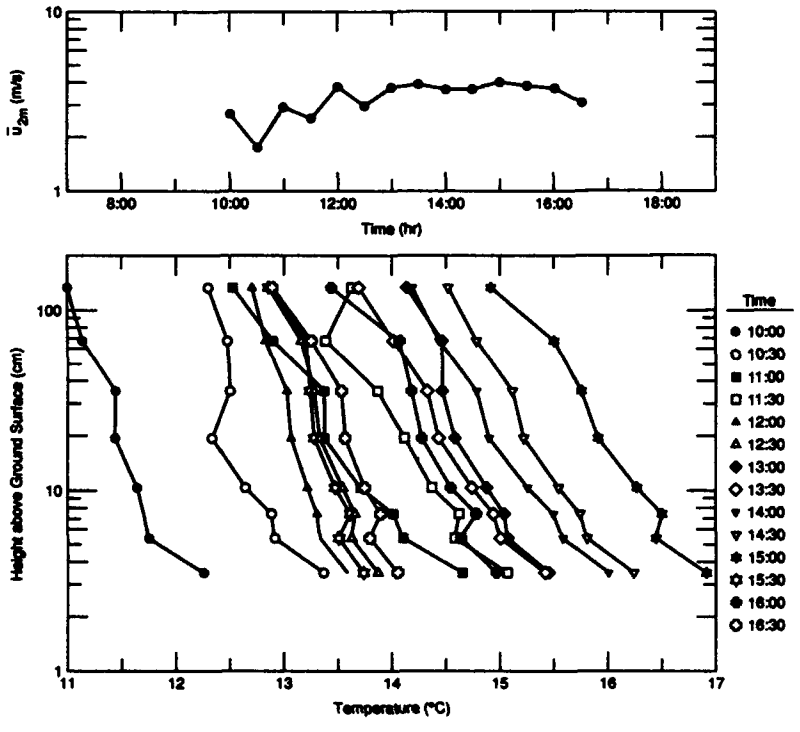


f. Over frozen ground.

Figure 2 (cont'd). Temperature and $\overline{U}_{2m}$ variations with height as function of time.

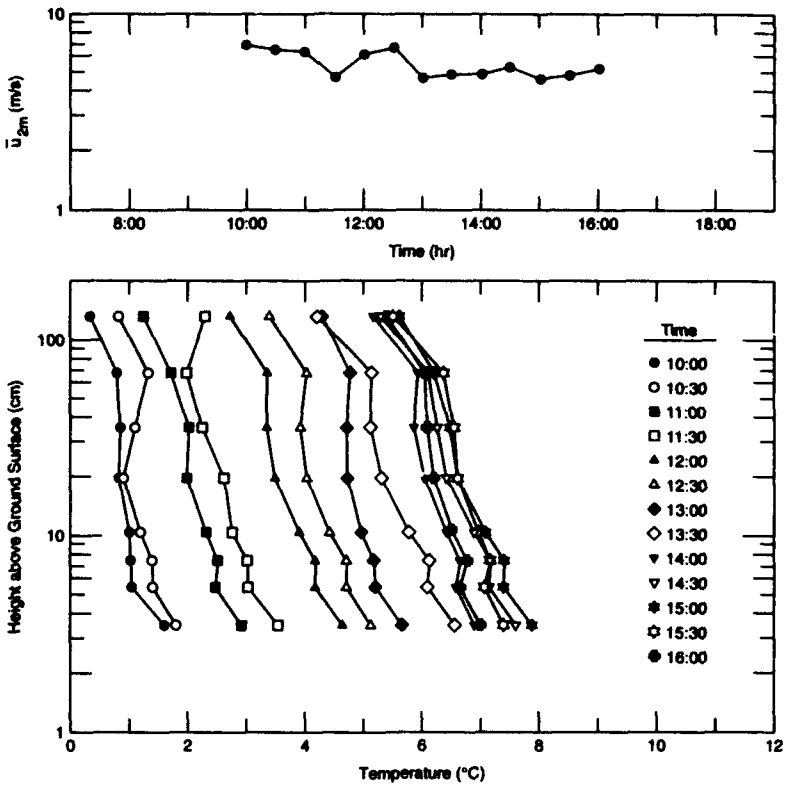


g. Over frozen ground.



h. Over partially thawed frozen ground.

Figure 2 (cont'd).



i. Over refrozen ground.

Figure 2 (cont'd). Temperature and $\overline{U}_{2m}$ variations with height as function of time.

frost in contact with the thermocouple beads (a uniform cylindrical shell of frost of approximately 2.0 cm in diameter has been observed on items at the site). Therefore, most of the sonic and temperature profile measurements were obtained from the mid-morning to the late afternoon, around or just before sunset.

Figure 2b shows another temperature profile over snow. In contrast to the profiles shown in Figure 2a, the temperature of the whole column of air was decreasing instead of increasing as time passed from 1030 to 1200. The wind speed $\overline{U}_{2m}$ increased linearly from about 2.0 to 4.7 m/s, but the temperature of the whole air profile decreased to a smaller extent from 1030 to 1130 and maintained a more or less identical profile shape, demonstrating distinctively a double-reversal in temperature (as shown in Fig. 2a). The wind speed was more or less constant, varying within the range of 4.0 to 4.7 m/s from 1200 to 1530, but the temperature profile at 1200 is drastically different from the others taken prior to 1200. The raised maximum temperature seemed to disappear and there were much greater variations in temperature as a function of height.

The solar radiation data determined from a nearby site (the data collected near the experimental site were incomplete, because not enough information was available to construct a diurnal radiation intensity distribution) also showed an unusual shape for this particular day. The solar intensity (net) started to decline at about 0840 from nearly 0 W/m^2 to a minimum value of approximately −120 W/m^2 (a sharp decline) at about 0940. This was followed by sharp reversal, and the solar intensity rapidly increased to a maximum of about 79 W/m^2. Afterwards, the solar intensity showed a normal decline until it reached a value of 0 W/m^2. The air temperature increased again and retained its regular shape, but with a less distinctive raised maximum temperature. At 1400 and afterwards, the temperatures started to decline again and began to form a regular and smooth temperature profile as shown in Figure 2a.

Figure 2c shows the temperature variation over hard-crusted snow. In general, it has almost identical features as shown in Figure 2a; namely, the temperature profiles have a similar distribution pattern, showing distinctly the double-reversal in temperature and progressively increasing with height from 1000 to 1400. From 1530 on, the whole temperature profile shifted lower, accompanied by the gradual disappearance of the double-reversal. The temperature profiles were also nearly

isothermal from 1 up to 128 cm at 2200, as well as at 0800 the next day. (Visual observation showed that there was no frost accumulation around the protruding thermocouple wires because there was cloud cover during the night). Once again, the profiles show that the double-reversal of the temperature structure was not caused by the characteristics of that particular thermocouple, but was caused by the effect of the water vapor near the surface layer with its ability to absorb solar radiation (proven because this double-reversal feature gradually reduced in intensity and disappeared completely at sunset). The net solar radiation intensity was 0 W/m^2 up to 0730 and rapidly increased to about 120 W/m^2 at around 0830. This was followed by a small-scale dip and its increases resumed again at 0940. A maximum intensity of 240 W/m^2 was reached at 1200, followed by a continuous but smooth reduction to 0 W/m^2 at 1630.

Figure 2d shows the variation of temperature over a thin layer of snow and ice (large snow crystals). The temperature profile was taken under a cloudless sky. Notice that the intensity of the temperature double-reversal has been reduced somewhat from the cases over snow cover as shown in Figures 2a–c; however, the shape of the temperature distribution is similar. Even from 1500 and afterwards, the double-reversal is still noticeable. There seems to be no correlation between the rate of temperature change (which can be obtained by dividing the total temperature change by the time elapsed at each height) and the wind speed $\overline{U}_{2m}$. It seems that the rate of temperature change is more or less related to the solar radiation intensity and its variation with time. However, there is an exception at 1430: Lower temperatures were recorded than at 1400, but at 1500, temperatures were recorded at all levels that were higher than those at 1400 and 1430. This clearly indicates that the atmosphere near the surface is simply too complex to make the delineation of the individual factors influencing the overall results possible.

Figure 2e shows the data taken when there was no snow or ice covering the ground (there was some snow between the short grass stems). The figure is nearly identical to Figure 2d and shows only a weak display of the double-reversal in the temperature structure. The rate of temperature change was almost identical for all the levels from 1030 to 1230 and then decreased to about half this value from 1230 to 1400. However, like Figure 2d, Figure 2e shows a much greater increase in the temperature change from 1430 to 1500, followed by lower temperatures and nearly isothermal conditions at 1530 and 1600.

Figure 2f shows the temperature variations accompanied by a considerable change of wind speed during the test period, which varied from approximately 1 to 3.5 m/s. There was no snow or ice over the frozen ground. The maximum rate of temperature change took place from 1030 to 1100 and 1300 to 1330, corresponding to the lowest wind speeds during the entire experimental period. On this day, the radiation intensity reached a maximum of 590 W/m^2 around 1000, dipped to 560 W/m^2 at 1100, further declined to about 400 W/m^2 at 1200, increased to about 590 W/m^2 at 1300, and then declined to 0 W/m^2 at 1630. The largest temperature increase occurred from 1300 to 1330. From 1230 to 1300, the temperature decreased at all heights with no change in the distribution pattern, while the wind speed decreased from 2.0 to 1.15 m/s. After reaching a maximum at all levels at 1330, the temperature decreased slowly at all heights but following the same pattern at 1400, 1430 and 1500. The double-reversal was noted at 1330, when the highest air temperature at all levels was reached.

Figure 2g shows the data taken over frozen ground only (no snow at all). There is no double-reversal; but, other than that, the temperature distribution shows the familiar form, i.e., it decreases as height increases. The rate of temperature change seems to be much smaller, even though the radiation intensities were higher than those for Figure 2f. As in Figure 2f, there is also a temperature reversal from higher to lower during 1130 to 1200, though the radiation intensity had its deepest dip at about 1100. On the other hand, the temperature rise was rather small from 1230 to 1330, even though the radiation intensity was at its peak of about 700 W/m^2. The radiation intensity was in steep decline, i.e., from 600 to 300 W/m^2, during 1330 to 1430, but the rate of temperature change was at its maximum value for the entire test period. This was followed by a much greater lowering of the air temperature from 1400 to 1430. It seems quite clear that there is no simple way to relate the temperature distribution to either the wind speed (maybe the wind direction as well, but we have no data for confirmation) or the radiation intensity.

Figure 2h shows the most atypical temperature distribution for this series of measurements. The radiation intensity had a number of peaks and valleys. The maximum radiation intensity of about 570 W/m^2 was reached at about 0930 and was followed by two large declines to approximately 250 W/m^2 at 1100 and 1420, with a small dip at 1300 to around 430 W/m^2. At the beginning of the experiment, i.e., from 1000 to 1300, $\overline{U}_{2m}$ varied from 1.7 to 4.0 m/s; afterwards, it was rather constant at about 4.0 m/s up to 1600, when it started to decline again. This may explain the irregular variation of the temperature readings from the thermocouple at the highest level at 1100 and 1130. The largest decline in temperature took place between 1130 and 1200, even though the radiation intensity was on the upswing. The

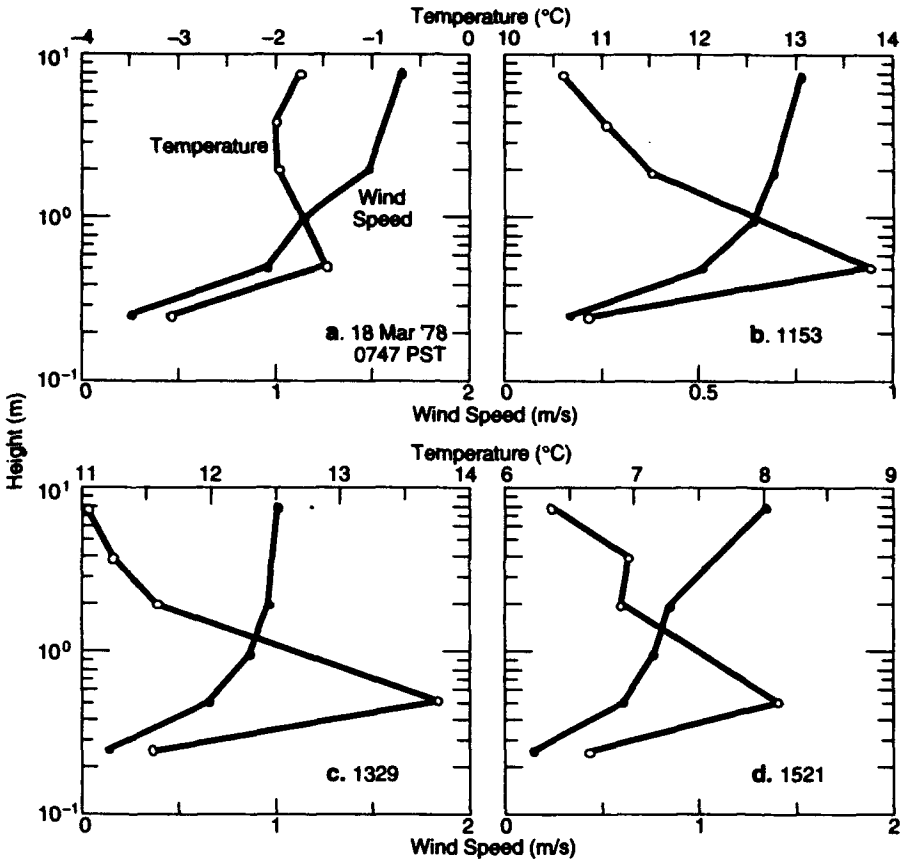


Figure 3. Temperature and wind speed variations with height over a melting snow surface (after Halberstam and Schieldge 1981).

largest temperature increase occurred from 1230 to 1300 followed by a slight decline at 1330; then, the temperature resumed its increase till 1500. There was nearly a 3.5°C decrease in temperature from 1500 to 1530, but the trend reversed itself and the temperature increased again from 1530 to 1600 (at the time, the solar intensity had decreased to 0 W/m^2). Therefore, it is impossible to explain the process of alternating increases and decreases in the air temperature during the period more or less affected solely by solar radiation, because, starting from 1300, $\overline{U}_{2m}$ is almost invariant at about 4.0 m/s. Contrary to the temperature profiles shown in Figures 2e–g, there were double-reversals in the temperature profiles at 1600 and 1630.

Figure 2i shows the temperature profiles taken over refrozen ground with no snow. It was a sunny day for the entire test period. Clearly, there was no noticeable double-reversal, with the exception at 1530 and 1600, in all the temperature profiles as shown in other graphs, except those for data taken over a continuous snow cover, patches of snow or a thin ice layer. The temperature was nearly isothermal between 5.5 and 7.5 cm. Also the temperature variations versus height appeared to follow a more regular pattern than those shown in Figure 2h. This may be attributable to the smoothness of the bell-shaped radiation intensity curve for this particular date, as opposed to that used for describing the radiation intensity characteristics for the data in Figure 2h. Similar to Figure 2h, there is a temperature reversal at 130.5 cm at 1130. Also, like the data shown in Figure 2h, the value of $\overline{U}_{2m}$ varied within 3 to 5 m/s from 1000 to 1300 and remained nearly constant after 1300. During most of the experimental period (daylight hours), the temperature rise over the frozen or partially frozen bare ground between the levels of 3.5 and 130.5 cm was in the range of 1 to 3°C. On the other hand, the temperature rise over a snow cover or a partial snow cover was much less and was only on the order of 0.1 to 1.5°C. Therefore, the overall temperature gradient over the bare ground was about twice that over snow-covered ground.

CONCLUSIONS

These limited and preliminary measurements of temperature profiles at 1 cm from the snow surface and 3.5 cm from the ground surface without snow at least partially confirm the results reported by Halberstam and Schieldge (1981). They reported, in their study of anomalous behavior of the atmospheric surface layer over a melting snowpack, a persistent raised maximum temperature at 50 cm above the melting snow surface. Though they measured the air temperature at 12 cm above the snow surface, no data were shown to ascertain whether the temperature decreases all the way from the raised maximum to the snow surface (at its melting temperature or lower). If this is the case, then there was no double-reversal in temperature near the surface layer (see Fig. 3). In my study, regardless of whether or not there was a snow cover, the air temperature always decreased first and then rose to a maximum and again decreased slowly as height increased. The raised maxima in this case are not as great as those reported by Halberstam and Schieldge. I believe that their much greater increase in the raised maximum is ascribable to the location of their field test site, which had about 2100 m of elevation and was fairly uniform in all directions for approximately 4 km, and experienced mostly low wind speeds during their experiments (see Fig. 3).

As pointed out by Halberstam and Schieldge, both the wind and the temperature rose and fell irregularly throughout their test period. The temperature profiles taken here over snow, or a mixture of snow and ice, surfaces show nearly isothermal conditions during most of my test period (excluding the period after the sunset). This is contrary to the results of Halberstam and Schieldge (1981); they found unstable conditions around the noon hour only (they found the temperature at 8 m to be lower than that at 25 cm during the noon hour but higher during rest of the period). They claimed that the profile below the maximum, on the other hand, was very stable, falling to the presumed 0°C surface temperature from 14°C in a distance of 0.5 m (see Fig. 3). However, based on this study, the temperatures at 1 cm (above a snow surface) and 3.5 cm (without snow cover) are, in general, always higher than the temperatures at 3.0 and 5.5 cm instead of being lower as pointed out by Halberstam and Schieldge. The temperature at 3.0 cm then increased to the raised maximum (in the case of a snow cover) at 5 cm above the surface. On the other hand, the air temperature from 5.5 to 7.5 cm (in the absence of a snow cover) neither increased nor decreased, indicating a thin isothermal air layer. From 7.5 cm on, the air temperature started to decrease slowly with height.

The existence of a thin isothermal layer has not been reported in the literature. The lack of discussion of this thermal feature may be attributed to the limited number of temperature devices installed near the snow surface by Halberstam and Schieldge (1981) and De La Casiniere (1974); thus, they were unable to delineate the true nature of the microthermal structure. There were no measurements between 0 and 25 cm above the snow surface in De La Casiniere's work, and the temperature distribution from about 75 cm upward was, in general, increasing instead of decreasing as those shown in Figure 2 of this study (see Fig. 4).

In a study related to energy exchange during the melting of a prairie snow cover, Granger (1977) reported two sequences of temperature profiles over melting snow taken in 10 April 1974 and 14 April 1975. In the 1974 measurement, the temperature profiles (see Fig. 5a) remained essentially stable through the day, indicating that energy was continuously being supplied to the snow surface from the air. There was no raised maximum and the increase of temperature was very small from 20 to 200 cm above the surface (i.e., nearly isothermal). On the other hand, the temperature profiles taken in 1975, although, like the 1974 measurements, stable at night, were unstable during the day (see Fig. 5b), as were those indicated in Figure 2 here. There were distinct raised maximum temperatures at about 20 cm above the melting surface, with a magnitude in the order of 2.5 to 3.5°C at 1200 and 1600 c.s.t. The temperature

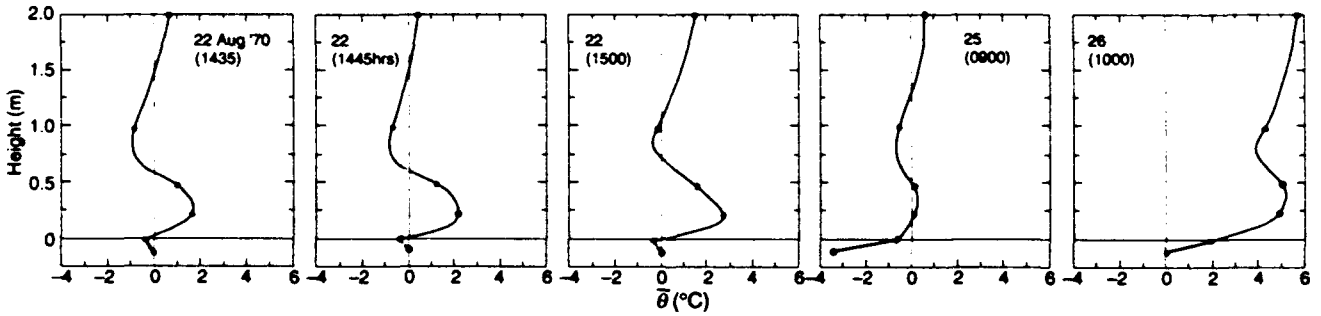


Figure 4. Temperature variation with height over a melting snow surface (after De La Casiniere 1974).

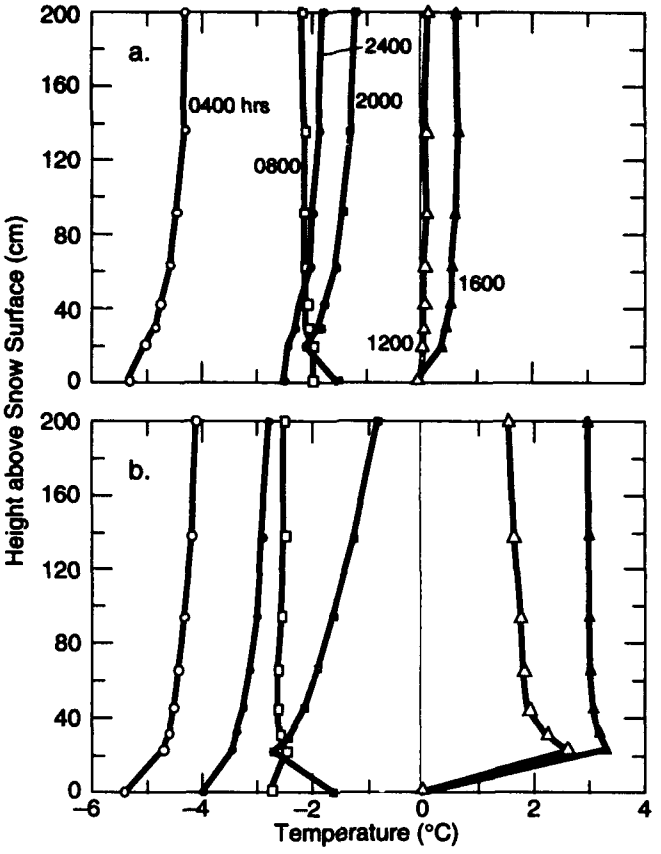


Figure 5. Temperature variations with height over a melting snow cover on a prairie (after Granger 1977).

then decreased gradually as height increased, similar to what happened in Figure 2. As in the works of Halberstam and Schieldge (1981) and De La Casiniere (1974), Granger did not measure the air temperature near the snow surface (in this case between 0 and 20 cm above the surface); thus, it is uncertain whether there was any inversion below the raised maximum. Granger (1977) further stated that the existence of unstable temperature profiles over snow when air temperature is higher than the ice melting point requires the presence of a divergence, or discontinuity, in the temperature field in the lowest centimeters of the boundary layer.

In this study, I observed a double-reversal in temperature structure for the case when both the temperature of snow and air were lower than the melting point of ice. The phenomenon seemed to occur as long as there was snow, regardless of how thick the layer was. Therefore, the divergence or the discontinuity in the temperature field existed even when both the air and snow temperature were lower than the melting point of ice. However, it seems that the temperature structure nearest the snow surface was much more complicated than those reported by Halberstam and Schieldge (1981), De La Casiniere (1974) and Granger (1977).

The work reported here provides a tentative confirmation of the temperature inversion feature in the air layer nearest the snow surface, regardless of whether the snow is melting or is lower than 0°C. Owing to the lack of snowfall and the lack of an appropriate field site, the data reported are very limited in scope. However, I am certain that the double-reversal in temperature structure near the surface layer was not attributable to defects in the temperature measuring devices. This was evidenced by the smooth temperature profiles exhibited at 1530 and 1600 (Fig. 2b). The double-reversal must be caused by intricate interactions of radiative absorption and emission by water vapor and turbulent convective energy exchange. To prove and to provide a concrete confirmation of this unique microthermal structure, further studies over a rather deep and well established snow cover will be conducted.

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Temperature profiles were taken during the 1991–92 season at an experimental site on CRREL's grounds; however, they were rather limited because of the unusual lack of snowfall and the location of the site, which was small and had elevated sides (especially in the direction of the prevailing wind). Nevertheless, some unusual features have been observed for the first time. The profiles show the existence of a double-reversal in temperature structure close to the snow surface when the snow temperature is much lower than its melting point, in contrast to a persistent raised maximum temperature over a melting snowpack. This indicates that the micro-heat transport process is complicated by the presence of a non-isothermal lower boundary. Over frozen or partially frozen ground without snow, and with air temperature either above or below the melting point of ice, a thin, nearly isothermal air layer on the order of 2 to 3 cm in thickness at varying heights was observed for the first time. With very few exceptions, in general, the temperature decreases as height increases, but the extent is less pronounced over a snow cover (where conditions are nearly isothermal) than over cold ground. On the basis of limited measurements, the overall temperature gradient over the bare ground is about twice that over snow-covered ground.

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