SOME OBSERVATIONS ON SNOWCOVER TEMPERATURE PATTERNS

Richard L. Armstrong
USDA Forest Service, Fort Collins, CO.

Introduction

The two primary boundaries of a snowcover are the snow-air interface (snow surface) and the snow-soil interface (soil surface). The soil surface is fixed, while the snow surface is a moving boundary. Temperatures found within a mountain snowcover normally range from -40°C to 0°C. Air temperatures may fluctuate well below 0°C, but soil temperatures beneath a snowcover are generally within a few degrees of 0°C. Therefore, temperature gradients at the macroscale are present to some extent unless conditions cause the snowcover to be isothermal at 0°C throughout. Gradients between the soil and the snow surface are largest in early winter, when the snowcover is relatively thin and air temperatures are low.

In his pioneering work on snow properties, Eugster (1952) recognized that a critical minimum temperature gradient must be exceeded before metamorphism in snow changes from the equitemperature (ET) to the temperature gradient (TG) mode (Sommerfeld and LaChapelle, 1970), but he did not specify the value of this minimum gradient, perhaps because he recognized that it varies over an appreciable range. Because the layers of a dry natural snowcover are always exposed to some degree of temperature gradient, the term "equitemperature metamorphism" is somewhat of a misnomer. Colbeck (in press) has addressed this problem specifically and describes two growth forms in the presence of a temperature gradient, a slower equilibrium form resulting in rounded grains and a more rapid kinetic form resulting in faceted crystals. Field workers often state that gradients in excess of 10°C/m are required to drive TG metamorphism. However, actual grain growth associated with TG metamorphism depends on the vapour pressure gradient resulting from the thermal gradient. Vapour concentration within the pore space of the ice matrix is dependent on the average temperature of the air in the pore space. Therefore, the magnitude of the vapour pressure gradient is controlled by the magnitude of the thermal gradient as well as by the average temperature of the thermal gradient.

This paper introduces the concept of a minimal vapour pressure gradient required to produce faceted TG grains rapidly enough to be important for the avalanche formation problems.

Heat sources for the gradients include not only the soil but also near-surface snow layers which have been warmed, primarily by solar radiation. This paper is an investigation of temperature conditions in the snow near the ground surface and at the snow surface. Near the ground surface, snow is subject to generally steady conditions which tend to alter snow texture over a period of weeks. At the snow surface, snow is subject to rapidly changing conditions which may alter the texture over a period of a few hours. Other work dealing with this subject includes Benson and Trabant (1972) and Bradley, Brown and Williams (1977), Marbouty (1981), and Armstrong (1981).

Heat Flow Through Snow

General heat flow conditions can be approximated by applying Fourier's equation for unidirectional steady state heat conduction in a homogeneous solid

$$\text{Heat Flow} = \lambda \frac{\Delta T}{\Delta Z} \tag{1}$$

which states that the rate of heat flow (W/m^2) across the thickness Z of a material is proportional to the vertical temperature gradient (°C/m) where the constant of proportionality λ is termed the thermal conductivity (W/m°C). However, because heat transfer in snow is not restricted to conduction alone, the term "effective conductivity" is often used. Yen (1969) contains examples of effective conductivity plotted as a function of density. In total, heat can be transferred in snow by the following processes: 1) conduction through the grains in contact; 2) conduction and convection through the void space or interstitial air; 3) radiant energy exchange via the void space; and 4) molecular vapour diffusion through the void space. In most cases, because the net radiation exchange between grains is small, radiant heat transfer is insignificant. The thermal conductivity of ice is about 100 times greater than that of air; therefore, the contribution of conduction through air is relatively unimportant. However, the contribution to overall heat flow by vapour diffusion can be significant, especially in low density snow. Heat transfer by convection can account for up to 1/3 of the effective conductivity, but results of studies remain theoretical (Akitaya, 1974; Palm and Tveitereid, 1979).

The Role of Vapour Pressure Gradient in TG Metamorphism

When field workers collect snow temperature data, they often look for the familiar 10°C/m, considered the minimal value necessary for rapid TG metamorphism. However, the primary mechanism responsible for the formation of TG grains is vapour transport resulting from the temperature gradient. While a temperature gradient must be present, it is the average temperature of the gradient which greatly influences the amount of water vapour available within the pore space of the snow layer.

Vapour flux (kg/m^2) is proportional to the gradient of the vapour concentration and can be described by

$$\text{Vapour Flux} \cong - \frac{D}{R\Delta Z} \frac{(P_1 - P_2)}{T_1 - T_2} \qquad (2)$$

where D is the diffusion coefficient dependent on atmospheric pressure and temperature, R is a gas constant of 460 J/kg°K, and P and T are pressure of water vapour and absolute temperature at the two levels within the snowcover. For a specific discussion of this process, see Perla (1978), and Colbeck (1981). For a current review of snowcover proprties in general, see Male (1980).

The 10°C/m cycle is significant when the average temperature is near 0°C; but as average temperatures decrease, the gradient must increase if the rate of crystal growth is to be maintained. Steep temperature gradients alone do not always provide adequate vapour pressure. This relationship is apparent in Figure 1, which demonstrates the response of vapour pressure to temperature gradient and average temperature. In studies by LaChapelle and Armstrong (1977) a vapour pressure gradient (VPG) of 5.0 mbar/m appeared to be the lower boundary for the rapid development of faceted crystals in low density (100-300 kg/m^3) snow. In this case the term "rapid" is intended to refer to a time scale appropriate to the problem of avalanche formation in a seasonal snowcover.

Figure 2 shows how the VPG would vary across hypothetical 1.0 m snow layers given three different surface temperatures and assuming a linear temperature gradient to 0°C at the soil surface. With the gradient of 35°C/m, a possibility during mid-winter at a sub-polar site, the average temperature near the surface is too low to allow a significant VPG to develop. The lower one-half of the snowcover is, however, warm enough for TG metamorphism. The -15°C example is typical for colder temperatures at a mid-latitude continental site. The warmer average temperatures produce a larger VPG in the near surface layers than in the sub-polar example, but the gentler temperature gradient limits the zone of significant VPG to the lowermost one-third of the snowcover. The -5°C example is appropriate for a mid-latitude maritime climate where average temperatures are closest to 0°C but temperature gradients are not adequate to promote a strong VPG.

Figure 3 shows the response of VPG to a fixed-surface temperature and varying snow thickness. It is appropriate to consider this type of relationship for a given site because a typical standard deviation for mean monthly air temperatures for early to mid-winter is generally less than 2.0°C, whereas standard deviations for snow thickness could well be 1.0 m or more. Variations caused by snow thickness, therefore, typically play a larger role in establishing vapour pressure gradients than do variations caused by average air temperature. It is apparent from Figure 3 that when the snow height reaches 1.0 m the opportunity for rapid recystallization is greatly diminished.

Conditions Near the Snow-Soil Interface

During the summer months, net heat flow into the soil is positive, and the amount of heat stored is significant. During fall, net heat flow reverses, and the stored heat is extracted as air temperatures diminish. When a snowcover develops, it acts as an insulating layer which inhibits heat flow from the soil to the air. Generally, the heat supplied by the soil to the snow-soil interface exceeds that which can be transferred across the snow layer to the colder snow-air interface. Because the snow-soil interface cannot be warmed much above 0°C, and less energy is required to raise the temperature of the soil than is required to melt snow, excess heat warms the ground. Heat transfer across the snow-soil interface decreases as the stored summer heat is depleted and as the snowcover increases in thickness. Figure 4 contains typical measured gradients for a high-altitude continental site in early winter during the coldest portion of the diurnal cycle. Under these conditions the upward heat flow (Watts/m^2) toward the snow-soil interface greatly exceeds the flow away from it. Throughout the winter season, the relative amount of heat in the soil varies but the flow remains in the same direction. Conditions at the snow-air interface are more complex as will be seen later in this paper.

The early winter conditions of warm ground and cold air temperatures combined with shallow, low-density snow provide temperature and vapour pressure gradients which promote rapid TG metamorphism. In order to monitor such conditions more closely, the average long-term temperature gradient across a snowcover can be calculated from general climatic data. The mean monthly air temperature divided by mean monthly snow thickness provides a calculated gradient. Figure 5 contains an example of calculated and measured gradients. In this example, the coefficient of determination (r^2) is 0.96. The average for a total of four examples was 0.95.

Snow-Air Interface

Conditions Near the Snow Surface

Snow depths greater than 1.0 m are accompanied by diminished temperature gradients in the lower snow layers and diminished influence of the ground heat source on TG metamorphism. The VPG requirements for rapid TG metamorphism could be approximated temporarily at other locations in the snowcover by way of the following events. A prolonged period of warm weather and/or significant amounts of rainfall could bring the entire snowcover to or near 0°C followed by an accumulation of new snow at lower temperatures and, finally, clear cold weather conditions. Adequate VPG values could exist at the base of the new snow layer influenced by the warmer old snow below and the colder snow and air temperature above. It is possible for relatively thick layers (0.1-0.4 m) of faceted TG grains to develop at a mid-pack location. Such conditions would be transient, with the temperature eventually returning to a more linear form. However, because of the capability of snow to store heat, the return to the linear temperature distribution may require several days. On a smaller scale, the same conditions could develop if a melt-freeze layer was buried by a cold snowfall before it had lost its stored heat.

Most of the diurnal temperature variations occurs within 0.25 m of the snow surface in dry, low density snow (100-300 kg/m^3). Figure 6 is an example of fluctuations typical of a mid-latitude continental site (level) with clear sky conditions. The temperature of the snow-air interface of a subfreezing snowcover can drop very rapidly at night under clear skies. This is caused primarily by the low thermal conductivity of snow which does not allow heat transfer from the warmer layers below to take place fast enough to compensate for the rapid radiation loss at the surface. Consequently, temperature gradients are very steep during the night and early morning, but, due to the low average temperature of the gradients, vapour pressures are low and faceted crystals do not develop as rapidly as might be expected. In Figure 6 the early morning VPG is approximatley 10 mbar/m. In the colder climates of the northern Rocky Mountains, recrystallization observed in surface layers may result from this process (Perla, 1981). During mid-day the vapour gradient is in the opposite direction and in Figure 6 exceeds 30 mbar/m. While these gradients may be steep, they are of short duration and are highly variable. In general, these diurnal fluctuations and reversals of vapour flux appear to produce a coarse structure and a slightly increasing grain size, but not the easily identifiable faceted crystals associated with unidirectional vapour flow.

The development of faceted crystals is possible in the near-steady-state conditions at the base of the layer of maximum diurnal fluctuation. Figure 7 is an example where minimum snow surface temperatures are between -25°C and -30°C and maximums are between 0°C and -5°C, a set of values quite possible at a continental site with clear sky conditions. Observations indicate that at a depth of about 0.25 m below the snow surface, these diurnal temperature fluctuations would be damped to a range between -12°C to -10°C. Within a 0.1 m layer at the base of the zone of maximum diurnal variation (1.20 m to 1.30 m in Figure 7), the temperature gradient would remain in the 30 to 40°/m range during the diurnal cycle with a VPG of from 5.9 to 6.3

mbar/m. Unlike the case in the previous example, the sign of the VPG does not change diurnally. If such conditions persisted for approximately one week, an identifiable layer of beginning TG snow should result.

Conditions at the Snow Surface

Crystal growth at the snow surface may be extremely rapid. One example of this condition is the formation of surface hoar. At night, the snow surface and air layer just a few mm above the snow surface cool rapidly in response to a negative radiation balance. Depending on temperature and humidity conditions, the air may be cooled below its dewpoint. Water vapour is then deposited as surface hoar crystals onto the snow surface. Surface hoar formation is most common at night under clear skies with little or no wind and when the air mass of the previous day has been warm and moist. Data from a dewpoint or relative humidity sensor, located in a standard weather instrument shelter some 1.5 m above the snow surface, would not adequately monitor the frequency of surface hoar formation. In general, under clear sky conditions, an inversion exists within the first few meters above the snow surface unless colder air is advected over the snow surface (McKay and Thurtell, 1978). Figure 8 provides an example of the typical temperature difference between air (recorded in a standard shelter) and the snow surface for a 24-hour period under clear skies. The temperature difference is greatest at night, often being 10°C or more. Typical daytime air humidity ranges are included with night-time values increasing proportional to the drop in temperature. In this example, the temperature at the snow-air interface has dropped below the dewpoint resulting in the formation of surface hoar. However, the temperature at the level of the shelter remains above the dewpoint throughout the night.

Snow can also recrystallize at the snow-air interface during the day. This example requires a closer look at the energy exchange conditions at the snow surface. While the major portion (70 to 90%) of the incoming short-wave solar radiation is reflected from a clean, dry snow surface, the unreflected portion is absorbed at the surface or penetrates a few centimeters and is absorbed within near-surface layers. Penetration of radiation is maximized when the rays of the sun are perpendicular to the snow surface and when snow density is low.

Theoretically, the amount of solar radiation available for absorption is maximum at the snow surface. However, in terms of the net energy balance, the location of greatest warming may not always be the actual surface. Most of the heat gained by absorption at the surface is lost to outgoing long-wave radiation when the snow surface is exposed to a clear cold sky. As the day progresses, the surface (layer A in Figure 9) does experience some heat gain but the subsurface layers (layer B in Figure 9) are gaining heat at a faster rate. This is possible because the subsurface layers are insulated from the intense long-wave radiation losses at the actual surface resulting in increased retention of the absorbed solar radiation. Steep temperature gradients develop due to this net difference in heat gain between layers A and B and may exceed 100°C/m with vapour pressure gradients of 50 to 100 mbar/m if temperatures are near the melting point. Under these conditions, faceted crystals have been observed by the author to develop within a few hours. If the heat gain in layer B is sufficient to cause melt, the resultant near-surface stratigraphy includes not only a layer of weak recrystallized grains to serve as a lubricating layer for a potential avalanche but also a melt-freeze crust as a sliding surface. This process appears to be the most common during spring on slopes with a southerly aspect, but it is possible throughout the winter on any slope receiving direct solar radiation at some time during the day. While this process, first described by LaChapelle (1970) as radiation recrystallization, results from a narrow range of radiation balance conditions, its occurrence is sufficiently frequent to be of interest to the avalanche forecaster.

Conclusions

When snow covers the ground, a unique set of circumstances exists at the snow-air interface compared with other natural surfaces. A high albedo with respect to short-wave solar energy and a high emissivity with respect to long-wave radiation result in low net radiation. Energy absorbed at a snow surface is only a small fraction of that which would be absorbed by a soil surface or vegetated surface. However, the unusual thermal and optical properties of snow and the fact that it is a crystalline material existing close to its melting point, mean even small energy fluctuations can cause significant structural changes. Low thermal conductivity and a porous texture enhance temperature and vapour pressure gradients. In early winter, vapour fluxes within snow layers near the warm ground produce faceted TG crystals over a period of a few weeks, while specific solar radiation balance conditions in near-surface layers can cause similar changes within a few hours.

Temperature patterns contorl snow metamorphism which in turn affects the physical, and eventually the mechanical, properties of the snowcover. Elastic properties are strongly dependent on intergranular bonding, while optical properties, such as albedo, depend on porosity and grain shape. Relationships among temperature, density, crystal size and shape, and the growth and decay of bonds are as yet imperfectly understood. However, when established thermal properties for ice and deterministic models of water vapour transfer are combined with careful snow temperature and texture measurements, the result is a better understanding of how layers of faceted crystals develop at various levels in the snowcover.

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Discussion

Hamre:

I want to point out that I found excellent correlations on Dick Armstrong's vapour pressure tables with conditions that were observed up north and also have documented cases of upper level temperature gradient development in the snowpack over a short period of time. If anybody is interested in that we have documentation on these effects.

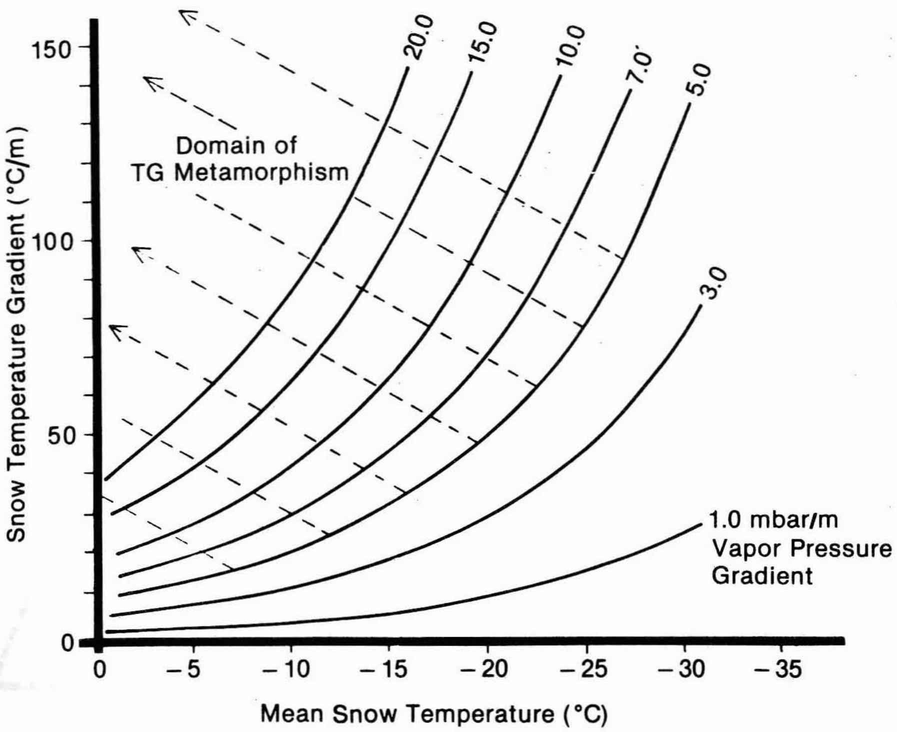

Figure 1. The relationship of vapor pressure gradient to temperature gradient and mean temperature within the pore space of a snowcover. The value of 5.0 mbar/m is considered the lower limit for TG metamorphism in low density snow.

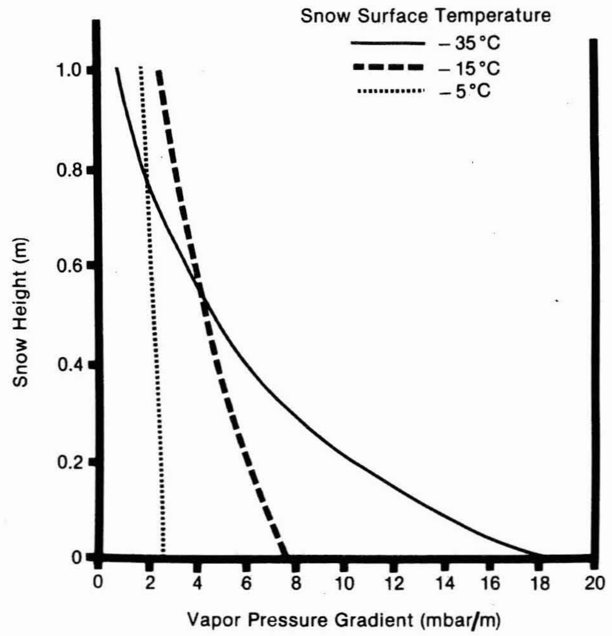

Figure 2. Variation in vapor pressure gradient across a 1.0 m thick snow layer resulting from three different surface temperatures (linear temperature gradients are assumed).

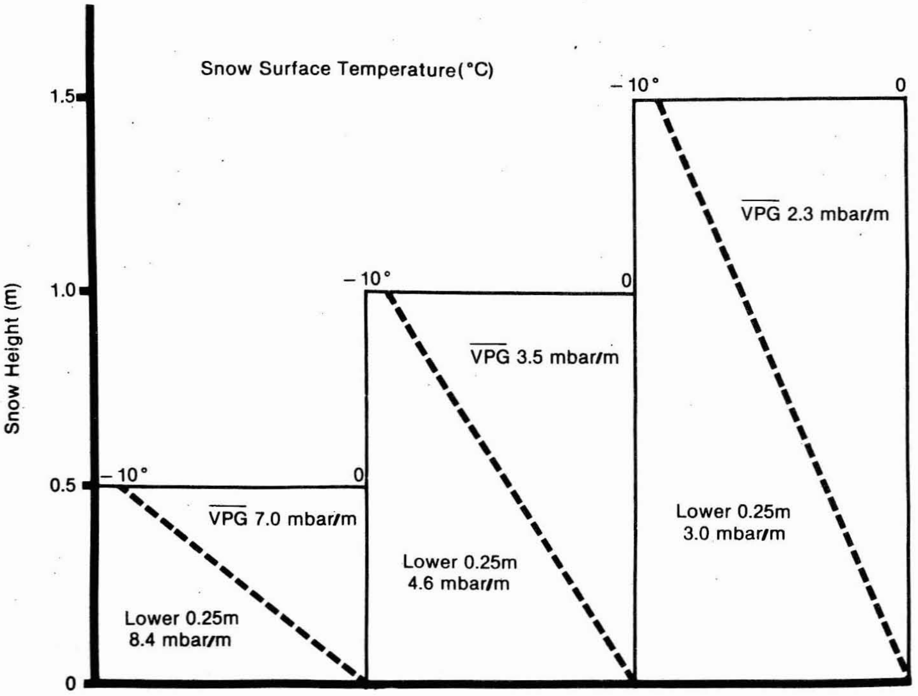

Figure 3. Variation in average vapor pressure gradient ($\overline{VPG}$) across the total snowcover, as well as VPG in the lower 0.25 m, with respect to three different snow heights and a fixed snow sueface temperature of -10°C. Dashed line is the "idealized" temperature gradient.

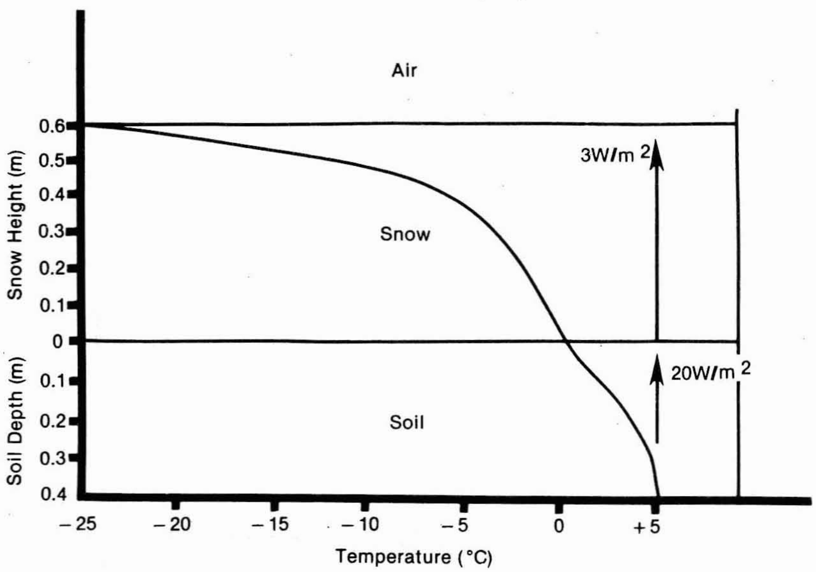

Figure 4. Temperature and associated heat flow conditions typical of a high-altitude continental site in early winter during the coldest portion of the diurnal cycle. (Red Mountain Pass, Colorado).

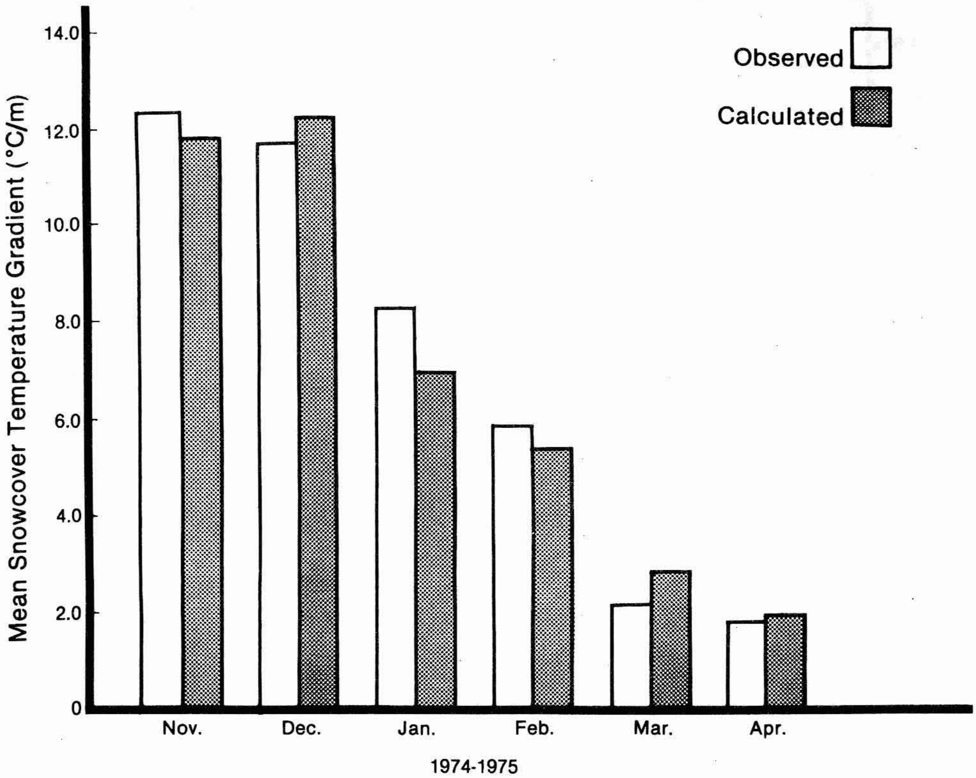

Figure 5. A comparison of observed and calculated mean monthly temperature gradients for the entire snowcover at Red Mountain Pass, Colorado, 1974-1975.

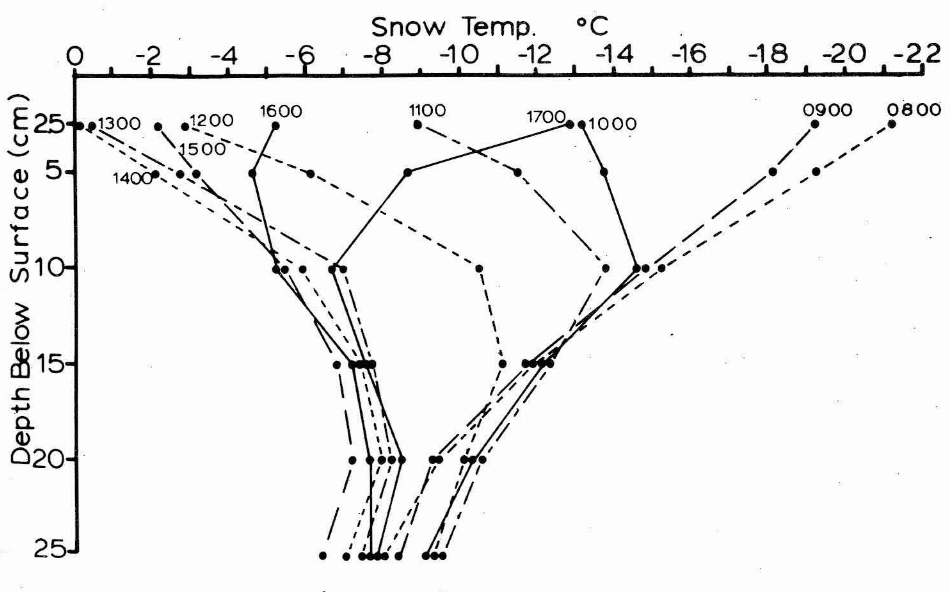

Figure 6. Hourly (0800 to 1700 LST) near-surface snow temperatures under clear skies at a level site, Red Mountain Pass, Colorado (3400 m), January 22, 1976. (average snow density 150 kg/m^3).

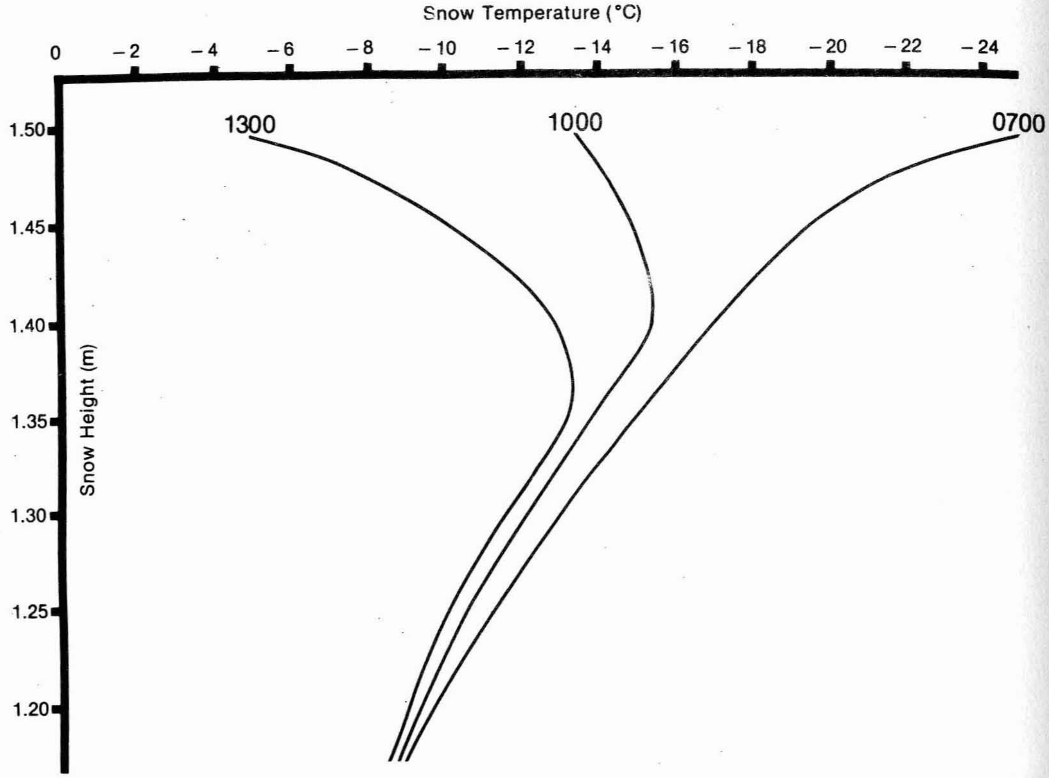

Figure 7. Idealized near-surface temperature gradients during a mid-winter cold spell for a high altitude continental site under clear skies. The 0700 and 1300 gradients represent the diurnal minimum and maximum values.

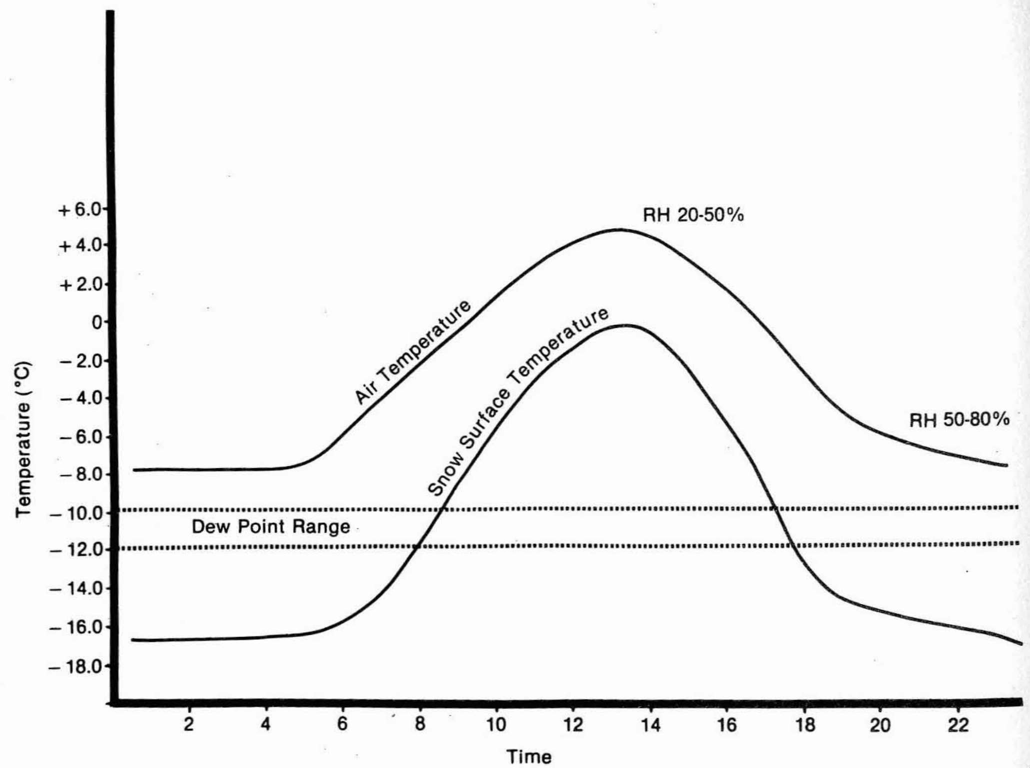

Figure 8. Clear-sky diurnal variation of air temperature (recorded in a standard shelter) and snow surface temperature (recorded by infrared thermometer) based on measurements during mid-winter at Red Mountain Pass, Colorado (3400 m). Typical ranges for daytime minimum and night-time maximum relative humidity at the shelter level are included with associated dewpoint range.

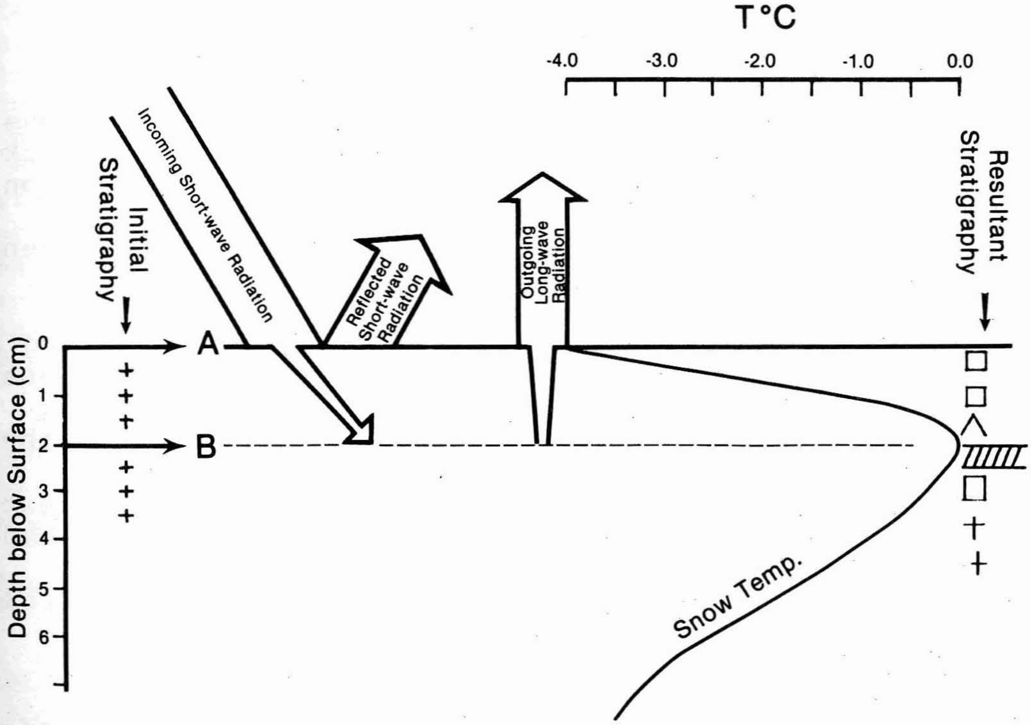

Figure 9. An example of radiation and heat balance conditions near the snow surface which would lead to the rapid recrystallization of the surface layer during the day. The temperature gradient is typical of those measured at mid-day during springtime under clear skies at a level study site in the San Juan Mountains, Colorado (elev. 3400 m).