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Boundary Layer Growth Over Snow and Soil Patches: Field Observations

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Abstract

Much of the snowmelt season is characterized by a patchy surface; differential heating of the snow and snow-free surfaces results in a significant horizontal transport of energy which affects and contributes to the snow melt. The calculation of the rate of energy advection requires some knowledge of the behavior of the thermal boundary layer over the patches of snow and snow-free surfaces. The paper presents the results from a series of field observations of the rate of growth of the thermal boundary layer over snow and snow-free patches. The results confirm that the boundary layer growth can be described by a power function of the distance from the leading edge of the patch. For the case of the thermal boundary layer over a snow patch within a bare field, the boundary layer growth is affected by the upwind surface roughness; the thermal boundary layer over a snow patch within a "rough" field grows much more quickly than that in a "smooth" field. Relationships are derived and presented for the parameterization of the boundary layer growth as a function of distance and upwind surface roughness.

Key Words: Boundary Layer, advection, snow melt, sensible heat, snow patches

Introduction

In the calculation of the melting of a patchy snow cover, the energy advected from the adjacent bare soil to the snow surface is an important consideration. As the lower layer air moves from the bare soil to the snow, it undergoes a considerable modification as energy is extracted. This additional flux of energy to the snow surface cannot be reliably calculated using traditional boundary layer flux-profile relationships, since these are based on the assumption of a constant flux layer. To date, only a very few attempts have been made to provide estimates of this advective flux over a field. Shook et al (1993) found that snow patch geometry and size frequency distributions could be described using fractal geometry. Synthetic snowcover spatial distributions were created using the fractal sum of pulses method and in a gridded calculation; an approximation of Weisman's (1977) model of advection over snow was applied to estimate the areal melt (Shook, 1995; Shook and Gray, 1997). Liston (1995) applied a numerical atmospheric boundary layer model, based on higher order turbulence closure assumptions, to simulate local advection during the melt of a hypothetical snow cover with simple geometrical properties; he used the model to demonstrate the increase in available melt energy as the snow cover is depleted. Essery (1999) also applied a modified boundary layer model to assess the advection of energy with varying snow cover fraction.

The boundary layer analyses described by Weisman (1977) and Liston (1995) require the solution of a very complex set of equations describing the conservation of momentum, mass and energy, as well as the changes in atmospheric stability as a boundary layer develops. Weisman reduced his results to a simple parametric form; however, he was only able to provide a few sample values of the coefficients required to apply his method. Weisman's approach assumes a fully developed boundary layer at the leading edge of the patch and, as such, cannot

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be applied with confidence to the early and middle stages of snow cover depletion when the surface consists of bare soil patches of short fetch.

The above references consider only isolated snow patches or patches distributed in a regular manner within a field. Application of any boundary layer analysis requires spatial knowledge of the patch sizes as well as the fetch distances. Snow patches display fractal attributes and self-similarity (Shook et al., 1993) and rarely align themselves in such a regular manner that calculations using uniform geometries are readily applicable to nature. Any successful method will necessarily require the incorporation of the statistical relationships describing the distribution of snow patch areas. Granger et al. (2002) showed that not only patch size distributions but patch fetch lengths could be described using fractal geometry and that as a result a hyperbolic frequency distribution of fetch lengths should develop for both snow and soil patches.

To deal with the complexity of changes in snow and soil patch geometry as snow coverage declines during melt, Marsh and Pomeroy (1996) developed the concept of advection efficiency, defined roughly as the fraction of the additional sensible heat produced over the bare patches that is transferred to the snow covered area. Advection efficiency is based on the premise that when bare soil fetches are short and snow fetches are relatively long (early melt) then almost all of the difference in sensible heat between bare soil and snow is advected to snow. However later in melt, when snow fetches are relatively short and bare fetches long, then not all of the heat difference over a domain will advect to snow, resulting in a declining 'advection efficiency' with declining snow covered area.

The advection efficiency approach represents an attempt to account for advection where relatively simple boundary layer analysis could not predict the advection term. However, it has remained a difficult method to use since the advection efficiency changes throughout the melt, and its application relies on the calibration of an empirical function, which can only be obtained through extensive field measurements of areal melt(Marsh et al., 1997). The concept remains appealingly simple for larger scale modeling (Cherkauer and Lettenmaier, 2003), but requires a physically-based method to calculate the advection of sensible heat from bare ground to snow as a function of the fetch lengths of both surfaces – so far a method to calculate this has not been developed.

Granger et al. (2002) developed a new approach, in which boundary layer integration is used to provide a means of calculating the amount of energy removed by the snow patch surface as warmer air moves over it. The method is based on the premise that in an advective situation, as in the case of a snow patch or a water body, the boundary layer at the leading edge of the surface change undergoes a significant modification as the relatively warm air moves over the colder surface; consequently, the amount of energy extracted to or from the surface can be obtained from the difference in the mean profiles at the leading and downstream edges of the patch.

Figure 1 is a schematic representation of a snow patch, showing the upwind and downwind conditions. As the air flows over the snow patch, an internal boundary layer develops, controlled by the difference in surface properties between the upwind bare surface and the snow surface. The surface roughness and the stability of the air control the height, B, to which the boundary layer will grow. Above the height, B, the "regional" boundary layer and vertical turbulent transfer of sensible heat, H, remain unaffected by the underlying snow. The temperature profiles below this level will be greatly altered (Figure 2). The downwind-moving air also carries energy in a horizontal direction; the transport of energy is simply the energy content of the air itself. As the air moves from a warm to a relatively cold surface, some of this horizontally transported energy is extracted to the cold surface.

The schematic representation of the temperature profiles is shown in Figure 2. The air immediately above the snow surface is stable (temperature increasing with height), while the regional boundary layer and the air over the bare surface are unstable. The temperature profile above the snow exhibits a maximum at or near the height B. Granger (1977), Male and Granger (1978) and Halberstam and Schieldge (1981) have documented the existence of such a temperature maximum in the lower boundary layer above a snow surface; these authors attributed the temperature maximum to radiative divergence.

Granger et al. (2002) showed that the advection of sensible heat to the snow patch surface is a function of the change in surface temperature between the bare soil and the snow, the change in vertical heat flux between the upwind and downwind edges of the patch, and the boundary layer height. The boundary layer height is, in turn, a function of the path length, or the size of the snow patch.

The boundary layer height, B, is in effect the height to which the internal boundary layer has grown as the air passes over the snow patch. Its behaviour and rate of growth under varying degrees of atmospheric stability has not been the focus of many studies. Brutsaert (1982) states that based on results from several studies, the thickness of the internal boundary layer can be approximated by a power function of fetch:

$$\delta = cX^b \tag{1}$$

in which δ is the thickness of the internal boundary layer at the distance X from the leading edge. For the full path length, X_s, over a snow patch, $\delta = B$, and Eq. 1 becomes:

$$B = c \cdot X_s^b \tag{2}$$

The coefficients c and b depend on stability, and c also depends on the upwind and downwind roughness heights. For neutral conditions, Brutsaert (1982) gives a value for c of 0.334. The exponent, b, increases with increasing instability; for neutral conditions, he suggests a value for b of 0.77. According to Panofsky (1973), for extremely unstable conditions b can be as large as 1.5. These studies reported values for the wind field within the turbulent boundary layer.

Field observations of the internal boundary layer growth over snow patches have been extremely rare. Takahara and Higuchi (1985) provided a very insightful experiment on the thermal modification of the air over melting snow; their study focused on the calculation of melt and, although they demonstrated the development of the internal boundary layer, they did not provide an analysis of its rate of growth. It is the purpose of this paper to test with field observations the hypotheses on the growth of the thermal boundary layer over snow patches that are inherent in the method of Granger et al. (2002) and to further consider the atmospheric conditions at height B and find values for the coefficients in Eq. 2. A companion paper (Essery et al. 2005) describes the modeling of the boundary layer and the advection of energy to snow patches.

Equipment and experimental procedure

Two portable instrument masts were constructed for the deployment of the boundary layer instrumentation. Each mast consisted of a rigid 2-m PVC tube, supported by a camera tripod; the tubes were marked with 1 cm graduations. A series of fine-wire type-E thermocouple air temperature sensors were constructed; these consisted of a male thermocouple connector, a thin 15-cm fiberglass or plastic tube to carry the thermocouple wire, and a 1.5-cm coated metal "fork" which held the thermocouple junction and exposed it to the ambient air. The female thermocouple connector was secured to a spring clamp which could be attached to the mast at any desired height. A battery-operated Campbell Scientific 21x data logger, which could also be attached to the tripod, received the wiring from all the thermocouple sensors. Each mast was thus equipped with 7 or 10 thermocouples, as well as with an infrared sensor (Apogee Instruments, with body temperature and emissivity corrections) to measure the surface temperature; one of the two masts also deployed a cup anemometer.

This equipment was extremely lightweight and easy to move about. With the thermocouple sensors deployed with an approximately logarithmic spacing between 1 cm and 180 cm, the arrangement allowed for the detailed simultaneous observation of the thermal boundary layers upstream and over a snow or soil patch. When a suitable snow or soil patch was identified; one mast was set up in a fixed upwind position, providing a measurement of the thermal boundary layer over field. The second mast was set up over the patch; the upwind distance to the edge of the patch was measured and the dataloggers were allowed to collect temperature and wind speed information for a period of 20 to 30 minutes, with 5-minute means. The mast was then moved to a new position, the fetch distance recorded, and data collected for another period. In this manner, a transect of observations was obtained, allowing for a determination of the rate of growth of the thermal boundary layer with distance from the leading edge of the patch. Figure 3 is a schematic drawing depicting the equipment set up over a snow patch.

Measurements were obtained for a variety of field situations, varying from agricultural fields to alpine tundra and short bush tundra. The agricultural fields, in central Saskatchewan, were generally very flat and smooth with large upwind fetch distances; when the snow melts on these fields, the exposed bare soil represents a strong thermal contrast with the remaining snow patches, but with little difference in surface roughness for the snow and bare surfaces. The alpine tundra and short bush tundra sites were located on an alpine plateau, in Wolf Creek Research Basin near Whitehorse, Yukon. In this area, snow melt exposes the vegetation first, resulting in a patchy surface for which there is a significant contrast in surface roughness between the exposed shrubby surface and the remaining snow surfaces. The study sites thus allowed for an examination of both the thermal contrast and the surface roughness change on the rate of boundary layer growth.

Results and Discussion

Figure 4a presents a typical temperature profile over a bare soil patch within a snow field, and Figure 4b presents the modified lower temperature profile over a snow patch(from the Saskatchewan data). Also shown in Fig. 4 are the upwind temperature profiles. The figures demonstrate how the thermal boundary layer is modified as it passes over a patch. In Fig. 4a we see that the temperature profile over a snow field is a stable one. As the air moves over a warmer bare patch, the lower boundary layer becomes heated through transfer of sensible heat from the surface to the air; the figure shows that at a distance of 7 m from the edge of a soil patch, the temperature profile has been modified to a height of approximately 30 cm. The internal boundary layer temperature profile over the bare soil is unstable; although it appears logarithmic near the surface, it shows an inflection with a minimum temperature near the boundary layer height.

Figure 4b shows a typical unstable temperature profile over a field of bare soil. As the air moves over a melting snow patch, sensible heat is transfer to the snow surface, cooling the lower air layer. The figure shows that, at a distance of 1.0 m from the leading edge of the snow patch, the thermal boundary layer has been modified to a height of approximately 30 cm. As with the soil patch, the temperature profile within the internal boundary layer appears logarithmic near the surface; it also shows an inflection with a maximum temperature, in the case shown, at a height of approximately 7 cm.

Figure 4 shows that the height of the boundary layer, *B*, can be estimated as the height at which the two temperature profiles effectively merge. In order to calculate this effective boundary layer height the difference between the upwind and downwind temperatures was plotted against the logarithm of height above the surface. A straight line was then fit through these points, and *B* was determined as the height at which the fitted difference approaches zero. Figure 5 presents the temperature differences for the profiles from Fig. 4; the best-fit line is an exponential in which the height of the internal boundary layer, *B*, is determined from the following equation:

$$Ht = B \cdot e^{c\Delta T} \tag{3}$$

in which *Ht* is the measurement height, and ΔT is the horizontal change in air temperature.

Fig 5a shows that over the bare soil patch, at a distance of 7.0m, the thermal boundary layer has developed to a height of 0.42m. For the snow patch, at a distance of 1.0 m from the leading edge, Fig 5b shows that the thermal boundary layer has developed to a height of 0.25m.

The boundary layer heights determined from Eq. 3 are then plotted against the fetch distance in Figure 6 for bare soil patches and in Figure 7 for snow patches. Fig. 6 shows data points for both a bare soil (agricultural fallow, 26 observations) and bare tundra (short alpine tundra, 11 observations). There are insufficient data to distinguish between the two situations. This might be expected since the upwind surface roughness appears to exert a controlling influence on the rate of boundary layer growth (see below). The best-fit line is a power function, given as:

$$B = 0.135 \cdot X^{0.6} \tag{4}$$

The power function is in agreement with Eq. 2 and with previously reported studies (Panofsky, 1973; Brutsaert , 1982), however, the coefficients are substantially different; Brutsaert (1982) had suggested a value of 0.77, and Panofsky (1973) stated that for extremely unstable conditions, the exponent can be as large as 1.5.

For the case of a snow patch within a bare field there exists the possibility of a greater range of upwind surface roughness. Fig. 7 shows the boundary layer height measurements over snow patches for three ranges of upwind surface roughness, representing approximately 125 observations. The surface roughness value was estimated from the observed vegetation heights using the formulations provided by Brutsaert (1983). The figure shows that indeed the upwind conditions do exert a significant influence on the rate of boundary layer growth; the thermal boundary layer over a snow patch within a "rough" field grows much more quickly than that in a "smooth" field. The best-fit lines in Fig. 6 are also power functions; these are given by:

for Zo=.002 – .005m $B = 0.18 \cdot X^{0.62}$ (5a)

for Zo=.02 – .04m $B = 0.23 \cdot X^{0.58}$ (5b)

for Zo=.08 – .12m $B = 0.64 \cdot X^{0.5}$ (5c)

In Eqs 5a,b,c, the exponents decrease approximately linearly with the upwind roughness, and the coefficients increase exponentially with *Zo*. The values for the exponents provided in Eqs 5 are much smaller than those reported previously in the literature for the wind field. However, they do correspond to the internal boundary layer height that can be inferred from the results of Takahara and Higuchi (1985). For the stated upwind roughness (0.006m), their figures 3 and 2 show that for fetches of 9m and 48 m, the internal boundary layer over snow has grown to heights of 0.6 and approximately 2.0m, respectively. These heights correspond closely to those predicted by Eq. 5a.

The effect of stability on the rate of growth of the thermal boundary layer is more difficult to determine. The relative importance of the "upwind" or the "downwind" stability might be expected to change for the case of a soil patch or a snow patch; the thermal boundary layer changes from stable to unstable as the air moves from a snow field to a bare soil patch; while the inverse occurs over a snow patch within a bare field. Weisman (1977) related the advected heat transfer to a dimensionless horizontal stability parameter A^*, given by:

$$A^* = -\frac{kgZ_0}{U_*^2}\frac{\left(T_{sX} - T_{su}\right)}{T_{su}} \tag{6}$$

in which k is the von Karman constant, g is the gravitational acceleration, U_* is the friction velocity, T_{sU} is the upwind surface temperature and T_{sX} is the surface temperature at distance X on the patch; the temperatures are expressed in Kelvin. In the study, infrared temperature sensors were used to measure the surface temperature and, with the 2m wind speed, it was possible to estimate the friction velocity and thus A^*. For one of the observation periods over a melting snow patch, the equipment was left stationary for an extended period of time, allowing the upwind and atmospheric temperature conditions to change. The boundary layer height, B, and the stability parameter, A^*, were calculated for each 5 minute observation interval; the results are presented in Figure 8.

Fig. 8 shows that, although there is some scatter in the data points, the boundary layer height does increase with an increase in the horizontal stability parameter. The best-fit line is given by $B = 1.23 + 1.34A^*$, with an r^2 value of 0.16. The range of conditions encountered in this single test was small; the effect, however, could be more significant as the diurnal cycling of the surface temperature of the bare surface can be large.

Summary

A pair of portable temperature profile masts was deployed to measure the growth of the thermal boundary layer over bare soil and snow patches. Results of the analysis confirm that the boundary layer growth can be described by a power function of the distance from the leading edge of the patch. (Eq. 2). For the case of the thermal boundary layer over a snow patch within a bare field, the boundary layer growth is affected by the upwind surface roughness; the thermal boundary layer over a snow patch within a "rough" field grows much more quickly than that in a "smooth" field.

An initial assessment of suggests that the rate of growth of the boundary layer may also be the affected by the change in stability from upwind to downwind conditions; the boundary layer height increases with an increase in the Weisman stability parameter, A^*.

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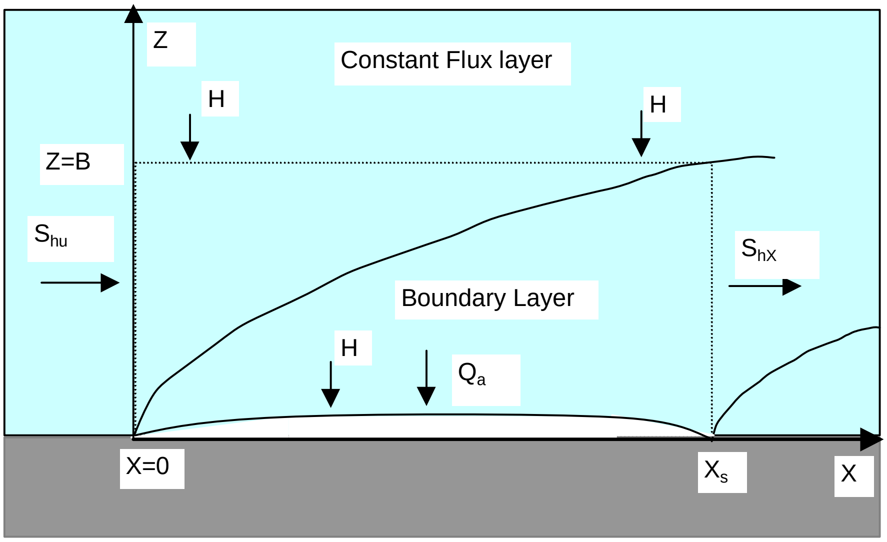

Figure 1. Schematic representation of the boundary layer development over a snow patch, showing the upwind and downwind energy conditions.

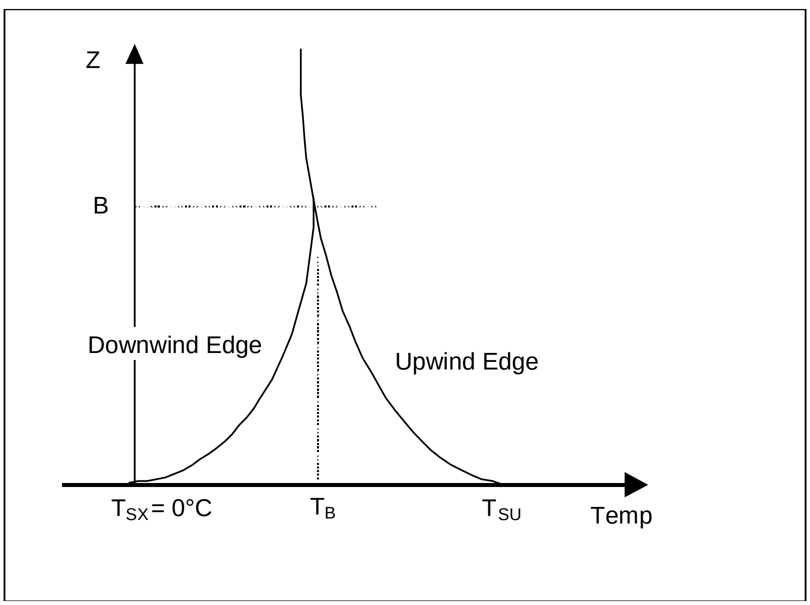

Figure 2. Schematic representation of the profiles of temperature at the upwind and downwind edges of a melting snow patch.

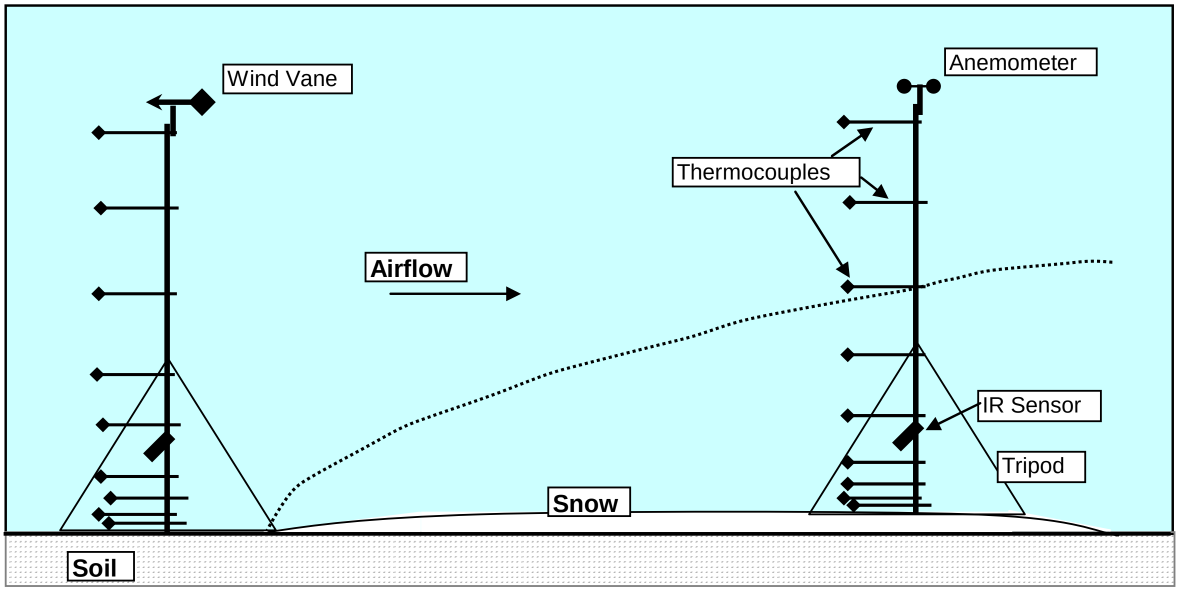

Figure 3. Schematic representation of the portable instrument masts set up at the upwind and downwind edges of a snow patch.

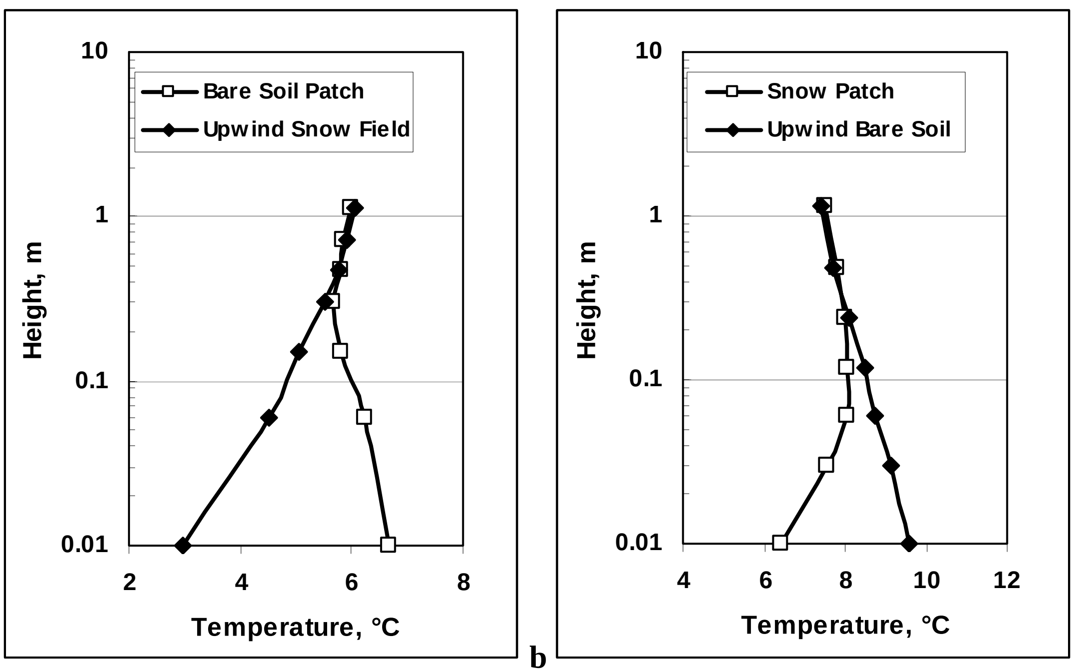

Figure 4. The temperature profiles (a) over a bare soil patch measured 7.0 m from the leading edge (2m wind = 6 m/s), with that over the upwind snow field, and (b) over a snow patch measured 1.0 m from the leading edge (2m wind = 6 m/s), with that over the upwind bare field.

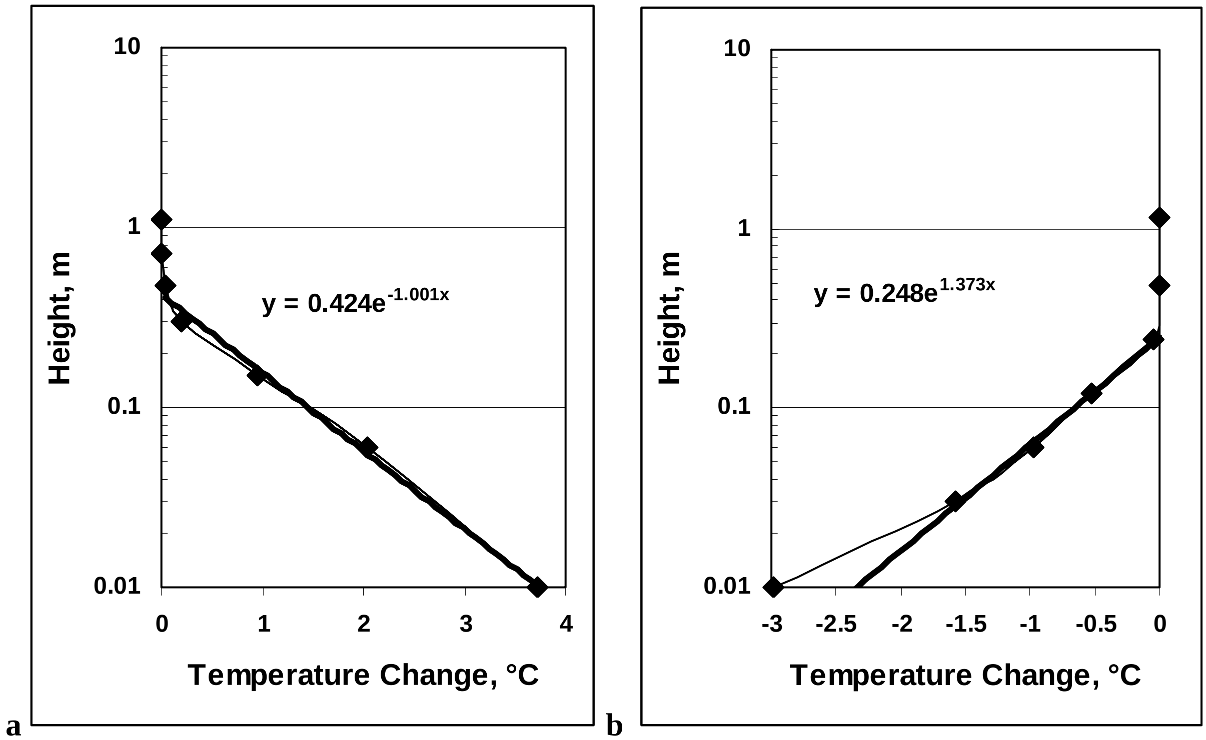

Figure 5. Plot of the temperature change (a) from a snow field to a bare soil patch, and (b) from a bare field to a snow patch. The best-fit lines provide the internal boundary layer heights.

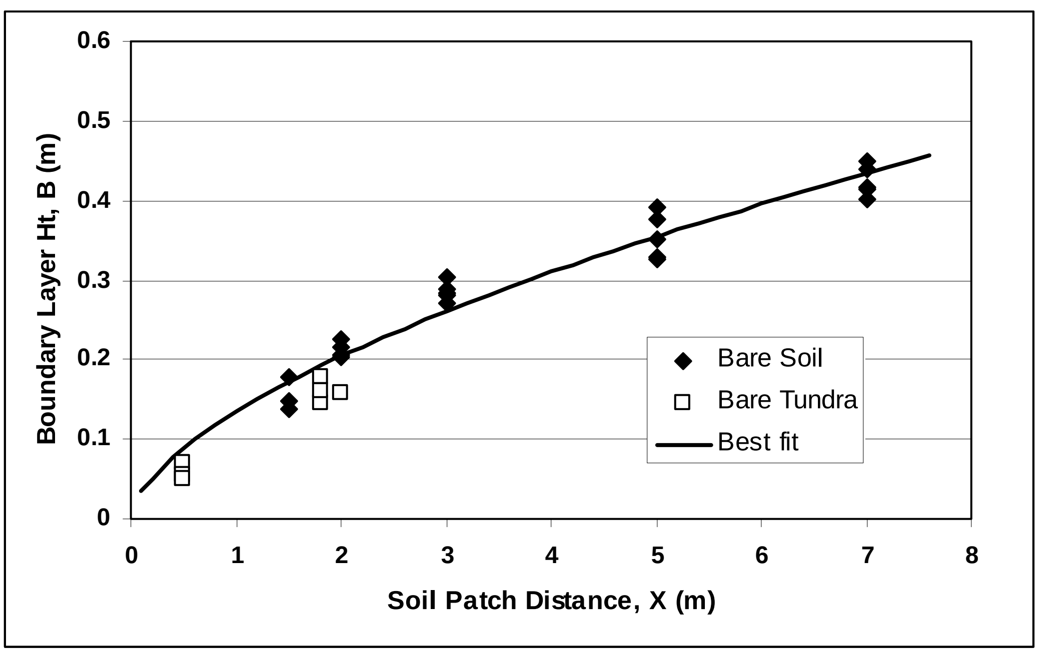

Figure 6. Thermal boundary layer height over bare ground patches within a snow field as a function of distance from the leading edge of the patch. The best-fit line is also shown.

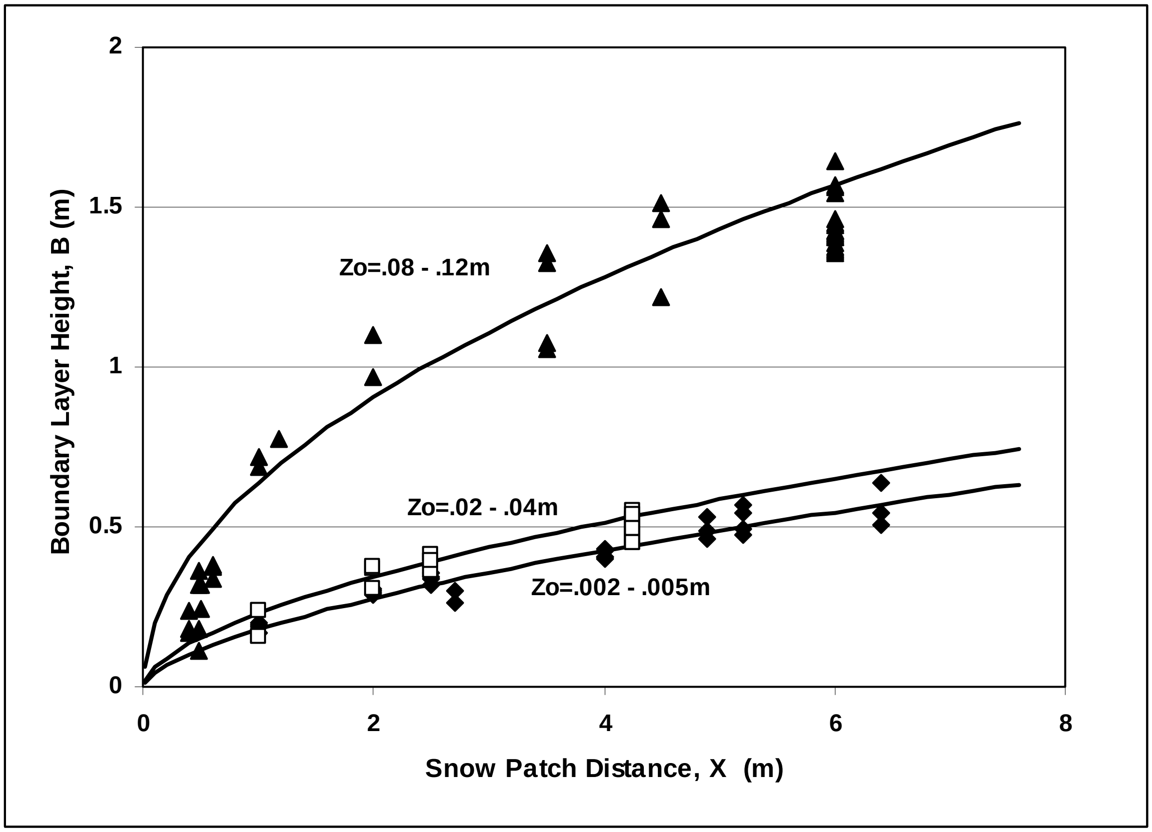

Figure 7. Thermal boundary layer height over snow patches within a bare field as a function of distance from the leading edge of the patch. The data are grouped according to the upwind surface roughness.

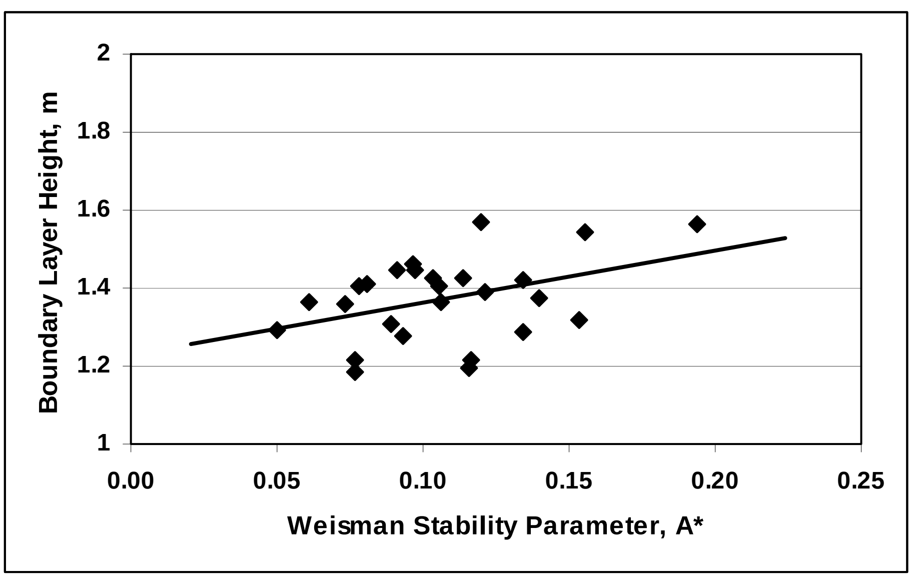

Figure 8. Plot of boundary layer height against the Weisman stability parameter, A*, for a melting snow patch. (X=6m)