DIFFERENCES BETWEEN AIR AND SNOW SURFACE TEMPERATURES DURING SNOW EVAPORATION

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ABSTRACT

When using the bulk aerodynamic formula for computing snow evaporation from the snowpack surface, investigators often only have access to air temperature measurements, and are forced to assume that snowpack surface temperature is the same as air temperature, at least for temperatures below 0°C. However, one of the main sources of energy for snow evaporation is air-to-snow sensible heat transfer. Such a transfer may require the snow to be substantially cooler that the overlying air. And measurements of snow surface temperatures do reveal temperatures often much colder than those of the air. In such situations, assuming equal air and snow surface temperature will result in the over-estimation of snow evaporation.

INTRODUCTION

Snow evaporation can be measured, estimated or computed in a variety of ways. The methods range from direct measurements of weight loss in snow pans to complex measurements of eddy fluxes using sophisticated technology. The most convenient computational method is the bulk aerodynamic approach (e.g. Moore, 1983). The convenience of this particular method lies in the fact that most of the required parameters can be obtained from regular meteorological stations. However, one of the necessary parameters is snow surface temperature. Snow surface temperature is troublesome for two reasons. First, it is not available from regular weather stations. Secondly, it is also rather hard to measure, a fact well known to anyone who has tried to obtain this value with thermistors or other physical-contact devices.

Because of the lack of proper data on snow surface temperature, it is not uncommon for researchers to assume that snow surface temperature is equal to air temperature or 0°C, whichever is lesser (e.g. Price and Dunne, 1976; Golding, 1978). Depending on the evaporative conditions, this assumption can lead to substantial errors in the estimation of snow evaporation. The purpose of this presentation is to examine the problems with the "equal temperature" assumption, and to determine the magnitude of the errors it creates.

THEORY

The snow surface gains or looses energy through many interrelated pathways: long wave and short wave radiative exchange, sensible heat exchange with the air, sensible heat exchange with the underlying snow, and latent heat exchange with the air. With the exception of short wave radiation input, all other forms of energy exchanges are tied to snow surface temperatures via various feedback mechanisms.

Snow evaporation,, or latent heat exchange from the snow to the air, is a process that lowers snow surface temperature. The driving force behind this exchange is the atmospheric demand for water vapour. The limiting factor is the speed at which energy can be supplied to the uppermost surface of the snowpack, and the speed at which water vapour can be transferred away from the snow surface. There are three possible sources of energy for snow evaporation: short wave radiation input, sensible heat transfer from the air and sensible heat transfer from the underlying pack. All three probably supply some of the energy required for the phase change. However, experimental evidence suggests that sensible heat transfer from the air and, under special conditions, from the underlying pack are the more direct suppliers of this energy. For example, during snow evaporation measurements carried out in Alberta by Bernier (in review), the largest evaporative events were recorded during evening or night periods, thus showing the relatively small importance of short wave radiation as a direct source of evaporative energy. In more direct measurements, McKay and Thurtell (1978) reported that measured fluxes in latent and sensible heat were often of equal magnitudes but of opposite signs over their snow pack, near Guelph, Ontario. The same pattern was observed more recently by Edwards _et al_. (1988) during a Ontario Hydro study on the deposition of atmospheric pollution. Some of the data from this study are shown in Figure 1.

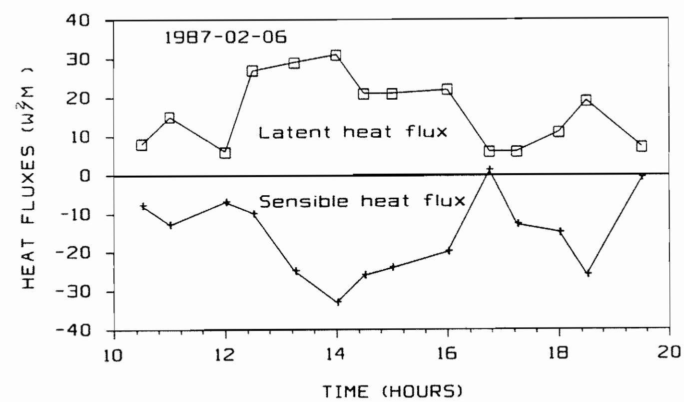

Figure 1: Latent and sensible heat fluxes over snow during an Ontario Hydro experiment on atmospheric pollution. Negative is towards the snow.

The linkage between latent and sensible heat is fairly straightforward. As evaporation proceeds, snow surface temperature is lowered until a sufficient temperature gradient exists to provide a downward sensible heat transfer that will satisfy the latent heat demand. At the same time, as the snow surface temperature decreases, the latent heat flux itself also decreases in response to the lowering in saturation vapour pressure at the surface of the snow grains. Of course, the latent and sensible heat fluxes are not always of equal magnitude as the other energy exchange processes also act as sources and sinks of energy by varying amounts. However, experimental data show that the link between sensible and latent heat flux is strong, and that the feedback between the two mechanism is expressed through modifications in snow surface temperature.

The current analysis uses the bulk aerodynamic equations as derived in Moore (1983). Sensible heat flux (Q_H) and latent heat flux (Q_E) can be expressed in a vertically integrated, or bulk, form as:

$$Q_H = \rho\, c_p\, D_H\, (T_a - T_s) \tag{1}$$

$$Q_E = \rho\, L_v\, D_E\, (q_a - q_s) \tag{2}$$

where ρ is the density of the air, c_p its specific heat, T_a and T_s the temperatures of the air ($z = a$) and of the snow surface, L_v the latent heat of vaporization (including fusion), q_a and q_s the specific humidity of the air at height a and of the air at the snow surface, D_H the bulk exchange coefficient for sensible heat, and D_E that for latent heat. Since both sensible and latent heat transfer are performed by the same eddy processes, it is generally assumed that $D_E = D_H$.

The bulk transfer coefficient is, of course, highly dependant on local wind and turbulence. However, over flat, open surfaces, logarithmic wind profiles develop, with regular and well defined vertical turbulent motions. Under such conditions, the bulk transfer coefficient can be defined as:

$$D_E = \frac{k^2 \, u_a}{\ln^2 (z_a/z_o)} \tag{3}$$

where: k is the von Karman constant (0.4), u_a is the wind speed at height z_a (cm s^{-1}), z_a is the height at which measurements are taken (cm), and z_o is the roughness length, as determined by measurements of the wind profile (cm) (see Moore, 1983, for typical values).

RESULTS

Figures 2 and 3 show the results of computations using the bulk aerodynamic equations shown above. Wind speed was set at 1 m s^{-1}, and roughness height at 0.4 cm. Wind speeds, air temperature and relative humidity were assumed to have been measured at z = 1 m. The computations were made in such a way as to have the fluxes of sensible and latent heat of equal magnitude, but of opposite sign, thereby letting snow surface temperatures settle at a level that accommodated both fluxes. According to the computations, snow surface temperature can be as much as 5°C to 8°C colder than air temperature, with the greatest differences occurring when rates of evaporation are highest.

In figure 2, we see that actual snow evaporation (latent heat flux transformed to mm of snow water equivalent evaporated per hour) varies with air temperature and relative humidity. The cross-over to condensation (negative evaporation) seen at air temperatures above 273°K comes about as snow surface temperature cannot increase above 273°K, with a fixed saturated water vapour pressure around 6 bars, while warmer air temperatures generate increasingly high water vapour pressures. As air temperature goes up, lower relative humidities are required in order to maintain a vapour pressure gradient towards the snow. As a result, warm air conditions are not very conducive to high evaporation rates.

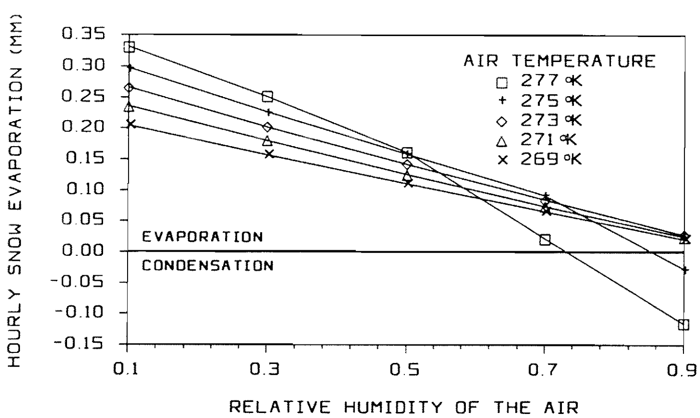

Figure 2. Computed snow evaporation for various air temperatures and relative humidities. Sensible and latent heat fluxes are assumed equal in magnitude, but opposite in sign.

Figure 3 shows the difference between snow evaporation rates obtained by assuming that air and snow temperatures are equal, up to 273°K, and those shown in figure 2. Once again, the 273°K barrier in snow surface temperature shows up quite graphically as the "equal temperature" assumption allows for a air-snow temperature gradient to develop once air temperatures exceed 273°K. As a result, the error in estimation gradually decreases with increasing air temperature as the postulated temperature gradient approaches the actual one.

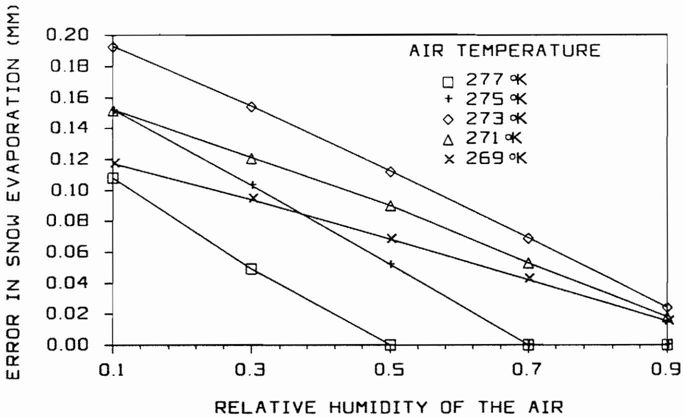

Figure 3. Difference between snow evaporation computed with the "equal temperature" assumption, and the values of figure 2, computed with a snow temperature based on a dynamic equilibrium between the sensible and latent heat fluxes.

CONCLUSION

Experimental results have shown that the sensible heat flux from the air to the snow supplies a large portion of the energy that escapes the snowpack surface through the latent heat flux. The results have also shown that snow surface temperatures are generally lower than air temperature as a result of the equilibrium reached between these and other energy exchange mechanisms. The above theoretical analysis has shown that latent heat flux can be overestimated if investigators assume that air and snow temperatures are equal when air temperatures are below 273°K. When air temperatures rise above freezing, the error in estimation decreases as the assumed air-snow temperature difference approaches the actual one.

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