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Evaporation from a Snow Cover

Review and Discussion of Measurements

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Possible evaporation rates from a snow surface with respect to available energy and vapour pressure deficit in the air are discussed. In literature reported measurements and measurements carried out at the University of Luleå are analysed. It is found that for northern areas the total amount of evaporation during the whole snow covered season amounts only to 10-20 mm.

Different formulas for estimating the evaporation from a snow surface are discussed. It is stressed that for forecasting purposes the aerodynamic formula and the profile method should be used only together with energy balance computations. The air density stratification reduces the evaporation rate considerably. A formula for the reduced evaporation rate is suggested. Finally, evaporation is analysed as a transient process, and possibilities of developing more soundly based evaporation models are discussed.

Introduction

It is, even among practising engineers, commonly thought that much water evaporates from a snow cover. This is an illusion. In spring the available heat produces little evaporation since the air is humid. In vinter, when the air is dry, the supply of heat to the snow is too small to vaporize it. Although hardly any water at all is lost from a snow cover through evaporation, the energy lost in this process might significantly influence the melting process. Also the converted process, sublimation or condensation, is from an energy point of view very significant.

Limiting Conditions for the Evaporation Process

For evaporation to take place from a snow surface two conditions must be fulfilled. Energy must be supplied to the snow and the vapour pressure at the snow surface must exceed that of the air. If the vapour pressure gradient is reverse condensation/ sublimation takes place. In order to be able to discuss the evaporation rate we must introduce the energy balance equation for a snow cover. It reads

$$R_s(1-a) + R_L - R_B + C - LE + G = \frac{dH_s}{dt} + sF \tag{1}$$

where R_s = incoming short wave radiation, a = albedo, R_L = incoming long wave radiation, R_B = from the snow cover outgoing long wave radiation, C = sensible heat transferred from the atmosphere to the snow cover, E = evaporation, L = latent heat of evaporation from snow, G = heat to the snow cover from the ground, H_s = heat stored in the snow cover, dH_s/dt = change of heat storage, s = melt rate, F = latent heat of fusion. The heat transport due to rain is omitted since it is of minor importance.

We must distinguish between melting and non-melting periods. During periods of melting the temperature of the snow is 0°C throughout the snow cover. Whence $dH_s/dt = 0$. During non-melting periods, of course, $sF = 0$.

The different terms in the energy balance is now estimated. The heat flow rate from the ground is small 2-4 cal/cm^2, day and can be neglected. The short wave radiation may vary from 0 to 500 cal/cm^2, day or more. The albedo is in the range 0.2-0.8. The long wave radiation is estimated from Stefan's law. The snow surface is almost a perfect black body but the emissivity of the atmosphere varies and can be as low as 0.7. Whence the net radiation may vary over a large range from say an input of 300 cal/cm^2, day to a net loss of 50 cal/cm^2, day.

The sensible heat flux, C, depends of course on the temperature difference between the snow surface and the air, but also on the stratification of the air and on the wind speed. Also the evaporation, which is due to a vapour pressure deficit in the air, is affected by the air stratification and the wind. Now we only have to remember, how differently the air temperature drops, when an air parcel raises dry - or wet adiabatically, or how convection currents generate above a warm surface, to realize that the radiation, the sensible heat flux, the evaporation rate, the wind speed close to the surface and the temperature stratification can't be evaluated separately, since they indeed affect each other.

During melting the radiation terms, the sensible heat transfer rate and the evaporation rate, e.g. the left hand side of the energy balance Eq. (1), do not depend on the terms on the right hand side, heat storage and melting. There is also a supply of energy for the evaporation process. However, during non-melting the snow surface temperature varies with time and may deviate from the air temperature. The back radiation and the heat fluxes depend on the surface temperature and on the heat storage.

In order to estimate the rate of change of heat storage let us assume the snow surface temperature to be constant at the air temperature and the temperature at the base of the snow cover to be T_g, which is also taken as the initial temperature throughout the snow cover. It is possible to find an analytical solution for the temperature profile in the snow cover and for the heat flux from the snow surface to the air. The heat flux, ϕ, is a Fourier – series

$$\phi = \rho_s ck \frac{T_g - T_a}{h} \{1 + 2 \sum_{n=1}^{\infty} \exp(-n^2 \pi^2 \frac{kt^2}{h})\} \quad (2)$$

where ρ_s = density of snow, c = specific heat of ice (half the value of that of water), K = the diffusivity of snow which varies depending on the density of the snow being 0.002 cm^2/sec for newly fallen snow and 0.004 cm^2/sec for old snow, h = depth of snow pack, t = time elapsed.

Taking T_g = 0°C, T_a = – 6°C and h = 100 cm we find for newly fallen snow that the heat flow from the snow cover after 1, 2, 12, 24 hours is 1.4, 1.0, 0.4, 0.3 cal/cm^2 h respectively. The heat loss during the first and the last twelve hours of the day is thus about 8 and 4 cal/cm^2. Whence, except for the very first hour after an air temperature change, the rate of change of stored heat cannot be a significant part of the energy balance equation. Therefore it is also clear that the snow surface temperature must adjust in such a way that thermal equilibrium is obtained.

Now, how much energy is available for evaporation during non-melting periods. Neglecting heat flow from the ground and heat storage we get the energy balance

$$C - LE - \sigma(273+T_s)^4 + \varepsilon\sigma(273+T_a)^4 + R_s(1-a) = 0 \quad (1b)$$

where σ = Stefan's constant, ε = atmosperhic emissivity

It is common to estimate the sensible and latent heat fluxes using formulas of the type

$$C = \text{funk}\,(W)\,(T_a - T_s) \quad (3)$$

$$LE = \text{funk}\,(W)\,(e_s - e_a) \quad (4)$$

where index s refers to snow surface and index a to atmosphere, W = wind speed and e = air vapour pressure. The wind function for the sensible heat flow is about 20 cal/cm^2/day/,°C and for the latent heat flow 30 cal/cm^2/day/mb. The temperature, T, is in °C.

Introducing the wind function for the sensible heat flow we can estimate the heat available for evaporation for some typical meteorological data (T_a = -10°C, ε = 0.85, a = 0.6, R_s = 100 cal/cm^2/day). From (1b) we now get approximately

$$LE = -330 - 30T_s \text{ cal/cm}^2\text{/day} \quad (1c)$$

Using Eq. (4) with the wind function 30 cal/cm^2/day/mb we get approximately in the temperature range around -10°C

$$LE = 153 - 78RH + 7.5T_s \quad (4b)$$

where RH = relative humidity of the air. Combining Eqs. (1c) and (4b) we find for RH = 100% T_s = -10.8°C and E = -0.1 mm/day and for RH = 70% T_s = -11.4°C and E = 0.2 mm/day. If the energy supply had not been limited, the surface temperature would not have dropped below the air temperature, and from Eq. (4b) the evaporation rate could then have been estimated at 0.0 (RH = 100%) and 0.7 mm/day (RH = 70%). The evaporation rate will, however, be even further reduced, since a stable air stratification develops, when the surface temperature drops.

During snowmelt there is of course energy available for evaporation. But even if energy is available there must be a vapour pressure deficit in the air. Is there? Say that the air temperature over a melting snow pack is 5°C and that the relative humidity is 0.7. The air vapour pressure is then 6.1 mb which corresponds exactly to the vapour pressure at the melting snow surface. Since the vapour pressure gradient is very small during the snow melt, evaporation as well as condensation takes place at a low rate. The evaporation rate during snow melt will be at maximum when the air temperature is just above 0°C. For high wind speeds the evaporation rate can from Eq. (3) be estimated at up to 0.5 mm/day.

The most favourable situation for evaporation is when a positive radiation balance exists, but still the air temperature is below 0°C. Also for this situation, however, the vapour pressure deficit hardly exceeds 2 mb, which means that the evaporation rate hardly exceeds 1 mm/24 hours.

Theories for Estimating the Evaporation

The evaporation from a snow cover can be determined in two ways – from profile measurements or from lysimeter readings. In the latter case the change of weight of a box full of snow is measured or snow pressure pillows are used. Some results obtained from these kind of measurements are discussed later.

From the theory of atmospheric turbulence, see for example Monin (1970), we know that when neutral conditions prevail, the logarithmic law describes the profiles of wind velocity, temperature and humidity in the air. We have

$$\frac{kW}{W_F} = \gamma \frac{T-T_0}{T^*} = \gamma_q \frac{q-q_0}{q^*} = f\left(\frac{z}{L^*}\right) - f\left(\frac{z_0}{L^*}\right) \tag{5}$$

where the function is

$$f\left(\frac{z}{L^*}\right) = \ln\left(\frac{z}{L^*}\right) \tag{6}$$

where z = height above snow cover, z_o = roughness length (where W = 0), W = wind speed at height z, T and q = temperature and humidity (weight over volume) at z, T_0 and q_o = temperature and humidity at z_o, k = von Kàrmáns konstant, W_F = friction velocity, γ and γ_q = ratio between exhange coefficients for momentum and heat, and momentum and mass respectively. T^*, q^*, L^* = temperature, humidity and length scales are defined as

$$T^* \equiv \frac{H}{k c_p \rho_a W_F} \tag{7}$$

$$q^* \equiv -\frac{\rho_w E}{k W_F} \tag{8}$$

$$L^* \equiv \frac{c_p \rho_a W_F^3 T_0}{\mathrm{kg}\, H} \tag{9}$$

where $H \equiv$ downward vertical sensible heat flux, $E =$ evaporation rate (upwards), $\rho_w =$ density of water, $\rho_a =$ density of air, $c_p =$ specific heat capacity of air at constant pressure and $g =$ acceleration of gravity. By measuring W, T, q at two levels it is possible to determine the scales and from them the evaporation rate and sensible heat flow rate.

When the wind is warm and the snow is cold as during melting the air is stably stratified. As long as Richardson's number $R_i < 0.15$, the logarithmic + linear law describes the profiles fairly well. Then

$$f\left(\frac{z}{L^*}\right) \equiv \ln\left(\frac{z}{L^*}\right) + \mathrm{const.}\left(\frac{z}{L^*}\right) \tag{10}$$

When the air is very stable the ratio between the buoyancy forces and the Reynold's stresses must not exceed a critical value. The profiles are then described by the function

$$f = 0.1 \frac{z}{L} \tag{11}$$

Using Eqs. (5) and (8) and the logarithmic law (6) the rate of evaporation can be estimated as

$$E = \rho_w^{-1} C_D W \gamma_q (q_0 - q) \tag{12a}$$

where the drag coefficient

$$C_D \equiv (k \ln^{-1}(\frac{z}{z_0}))^2 \tag{13}$$

Now since approximately

$$q \equiv \rho_a \frac{5}{8} \frac{e}{p} \tag{14}$$

where $e =$ air vapour pressure and $p =$ air pressure, Eq. (12a) is

$$E = \frac{\rho_a}{\rho_w} \frac{5}{8} C_D W \gamma_q \frac{(e_0 - e)}{p} \tag{12b}$$

or approximately in *SI*-units

$$E \equiv 8 \cdot 10^{-7} \gamma_q C_D W (e_0 - e) \tag{12c}$$

For neutral conditions γ_q is close to unity and the drag coefficient related to a height of 2 meters about $2\text{-}3 \cdot 10^{-3}$. Bengtsson (1975) found $z_o = 0.02$ cm corresponding to $C_D = 2 \cdot 10^{-3}$. Measurements carried out in the Soviet Union by Kuzmin, Kopanev and Ogneva reported in Konstantinov (1966) show that $z_o = 0.05$ cm and $C_D = 2.3 \times 10^{-3}$. From other measurements Konstantinov finds $z_o = 0.12$ cm and $C_D = 2.9 \times 10^{-3}$.

We can use the profile method for measuring the evaporation rate, but in the form so far used it is not very well applicable for forecasting evaporation. We need a formula from which we can estimate the evaporation with wind speed, air temperature, air humidity at a fairly large height and surface conditions as input. The temperature of the surface itself is also measured when performing gradient measurements. We know that the temperature T_o at z_o can be determined from an analysis of wind and temperature profiles. The ratio

$$m = \frac{T_0 - T_{2.0}}{T_s - T_{2.0}} \tag{15}$$

using s as index for surface can of course also be determined. If z_o is constant then m should depend on the temperature stratification of the air. However, if z_o depends on the stratification, there need not be a relation between m and the stratification. Numerous russian experiments (Konstantinov 1966) show that $m = 0.5$ over snow. This ratio also holds for the corresponding humidities.

Inserting $m = 0.5$, $\gamma_q = 1$ and $C_D = 2.3 \cdot 10^{-3}$ into Eq. (12c) we get in SI-units

$$E = 9.2 \cdot 10^{-10} W (e_s - e) \tag{12d}$$

or

$$E = 0.08 \, W(e_s - e) \text{ mm/day} \tag{12e}$$

with W in m/sec and $(e_s - e)$ in mb.

Applying the concept of Bowen ratio to an empirical formula for sensible heat transfer frequently used by Canadian researchers, Principles of Hydrology (1973), one can derive a formula of the form (12e) with a coefficient of 0.10 mm/day/mb/m/sec. Using the log+lin-law we will get a much more complicated expression such that

$$E = \Lambda E_n = \frac{\gamma_q}{\gamma_{qn}} \Lambda_1 E_n \tag{16}$$

where E_n = evaporation at neutral stratification, Λ_1 and Λ = stability factors. Index n for the ratio coefficient refers to neutral conditions. Using Eqs. (5), (8), (15) and the log + linear formula (10) one finds

$$\Lambda_1' = 1 - 2 \text{ const } \frac{m \gamma g z (T_a - T_s)}{T_a W^2} \tag{17}$$

where const is the constant in the log+linear expression, T_a = temperature in K^0, γ = Prandtl number as in Eq. (5) and $z = 2$ m is the height at which the wind speed and the temperature are measured. The correction factor is, putting γ const = 3, which was suggested by Deacon (1962), and introducing a Richardson number, Ri, evaluated between the surface and height 2 m, approximately

$$\Lambda_1 = 1 - 3 Ri \tag{18}$$

When introducing a correction factor one should also take into account that the ratio γ_q is reduced when the air is stable. Also the roughness length, z_o, changes (increases) slightly but this is of minor importance. When the air is so stable that the mechanical turbulence is surpressed $\gamma_q = 0.1$ and $Ri = 1$, Monin (1970). Since γ_q depends on the stratification and thus on Richardson's number the correction factor can be put on the form

$$\Lambda = (1-a_1 \cdot Ri)^b \tag{19}$$

Where a_1, b are coefficients.

From data given by Konstantinov (1966, p154) it is found that

$$\frac{\gamma_q}{\gamma_{qn}} \equiv \frac{1+0.72\,(\sqrt{1-cRi} - 1)}{1+0.30\,(\sqrt{1-cRi} - 1)} \tag{20}$$

The constant c is for snow about 2.5. For small values of Ri we get approximately

$$\gamma_q = 1 - 0.5\,Ri \tag{21}$$

and consequently for near-neutral conditions putting b in Eq. (19) = 1 and γ (neutral) = 1

$$\Lambda \equiv \Lambda_1\,\gamma_q \equiv 1 - 3.5\,Ri \tag{22}$$

A similiar correction factor for the effect of stratification on the vertical heat transfer is suggested by Monteith (1957). He used

$$\Lambda = (1+10\,Ri)^{-1} \tag{23}$$

in which the coefficient should be reduced by a factor 2, if the Richardson number is evaluated between the snow surface and height 2 m.

The influence of the correction factor is small for high wind speeds. Taking measurements at 2 m height to be $W = 1$ and 5 m/sec respectively and $T = 3°C$ we find that over a melting snow surface $Ri = 0.22$ (0.009), which gives $\Lambda = 0.23$ for $W = 1$ m/sec and $\Lambda = 0.97$ for $W = 5$ m/sec. It is also clear that for low wind speeds a stable stratification effectively surpresses the exchange of water vapour. The different evaporation equations are compared for three situations in Table 1.

Table 1 - Evaporation rate (mm/day) estimated from the profile method and the aerodynamic formula when $T_a \equiv 3^0C$, $T_s \equiv 0°C$, $RH \equiv 70\%$

Formula	$W = 1$ m/sec	$W = 5$ m/sec	$W = 10$ m/sec
aerodynamic			
$a = 0.18$, $b = 0.54$	0.22	0.53	1.15
$a = 0.13$. $b = 0.72$	0.18	0.47	1.07
profile Eq. (12e)	0.06	0.32	0.63
Eq. (22) stability included	0.01	0.31	0.63

It should be stressed that an aerodynamic formula is supposed to be used at least at a daily basis, whereas the profile method should be used on a much shorter time basis.

Variation over the Day

The air temperature and the energy supply to the snow cover varies over the day. Since there is no linear correlation between temperature and the saturated vapour pressure, and since both evaporation and condensation can take place during the same day and also snow melt can take place, it is hard to estimate the total evaporation and the total condensation and also the net moist transfer unless data are evaluated on a short time basis.

The variation of the air temperature and the vapour pressure deficit of the air over the day April 24 1979 at Luleå is shown in Fig. 1. Bengtsson and Westerström (1979) calculated the latent and sensible heat transfer rates as shown in Fig. 2. The exchange coefficient hypotethis was used and the effect of stable stratification was taken into account in the way suggested by Monteith. The drag coefficient was estimated at 2.8×10^{-3}. Since the net radiation and the snow melt rate were measured, and calculated and observed melt rate agreed well, the calculated heat flow rates should be quite accurate.

Using the aerodynamic formula with a = 0.13 mm/day/mb and b = 0.72 sec/m Lundberg (1979) found the latent heat flow to vary over the day of April 6 1978 as shown in Fig. 3. The maximum rate corresponds to 0.9 mm/day, but the calculated net total evaporation is using this method found to be less than 0.5 mm. However, during the night no energy is available for the evaporation process. The calculated sensible heat flow and the measured net radiation are shown in Fig. 4. Whence, after hours 21 no evaporation can take place. The net total evaporation is, when the energy supply is accounted for, calculated to only 0.14 mm.

Variation over the Winter Season

In mid-winter very little solar energy is available so even if the air is dry evaporation can hardly take place. In the spring radiation energy is available, but then there is only a very small vapour pressure deficit in the atmosphere. At the end of the melting period the vapour pressure of the air may be in excess of that of melting snow so that condensation takes place. The daily evaporation/condensation rates during the snow melt period in Luleå 1979 are shown in Fig. 5. During the last week condensation outweighted evaporation. The total evaporation was 6 mm but the net evaporation (evaporation-condensation) was only 3 mm. For an open field the total evaporation over the entire winter season was estimated to 17 mm. Hardly any condensation took place.

Latent heat transfer from or to a snow cover can be studied at glaciers also during summer time. For the Kårsa Glacier in Sweden Wallén (48/49) found that evaporation took place in May but during the summer vapour condensated at the snow surface.

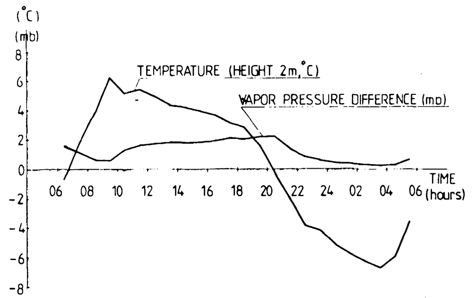

Fig. 1. Vapor pressure difference (e_s-e) between the snow surface and the air and temperature at height 2 above snow surface in a study plot of WREL, April 24, 1979.

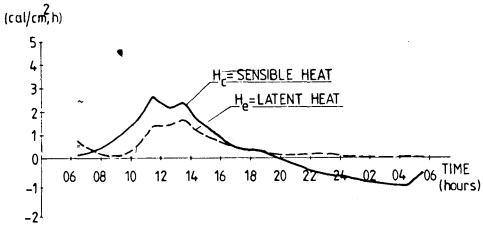

Fig. 2. Calculated latent (broken line) and sensible (unbroken line) heat flux to the snow pack for the situation shown in Fig. 1.

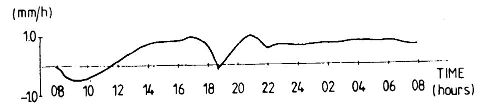

Fig. 3. Calculated evaporation (aerodynamic formula) from a snow covered open field at Bensbyn, Luleå, April 6, 1978.

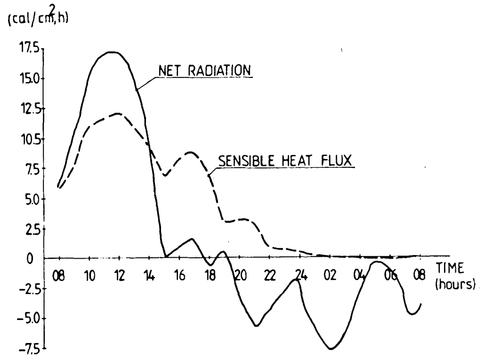

Fig. 4. Calculated sensible heat flux (aerodynamic formula) and measured net radiation to a snow covered open field at Bensbyn, Luleå, April 6, 1978.

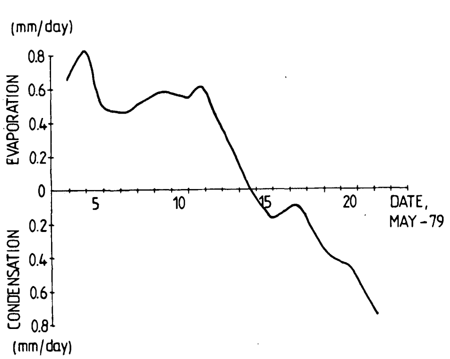

Fig. 5. Calculated evaporation rates from a snow covered forest site during the melting period 1979 at Bensbyn, Luleå.

Some Results from the Literature

Already Sverdrup (1936) found that turbulence may overcome the stability of warm air overlying snow at 0°C. West and Knoerr (1959) measured evaporation from snow cover in the Sierra Nevada. In a small opening in the forest they measured the evaporation loss over the winter season to 22 mm and in the forest itself to 8 mm. Much of the total occurred in one period of winds from the ocean. Toward the end of the season condensation on the snow outweighted evaporation. Later West (1962) calculated 3 year means for the opening to 33 mm and 16 mm in the forest site.

In Colorado Bergen and Swanson (1964) found daytime evaporation during sunny days in early spring to average about 1 mm.

In Michigan Treidl (1970) measured the loss of energy during 9 hours in February. The latent heat flux to the snow was 5 cal/cm^2, h, so that the condensation rate was 0.09 mm/h making a total of 0.78 mm.

For the Kårsa Glacier Wallen (1949) found that condensation of vapour supplied about 10% of the energy used in melting snow. The mean condensation rate for the period May-September 1930-39 was 0.23 mm/day. In favourable weather the amount of water condensed onto the snow was 0.7 mm/day.

Konstantinov (1966) summarizes some russian measurements. During snowmelt in April the mean evaporation rate for 27 days was 0.8 mm/day and during non-melting periods in March and April 0.4 mm/day. Konstantinov compares evaporation rates measured in lysimeters and rates calculated using the profile method and obtains good agreement.

Water balance computations for the Lappträsket representative basin in Sweden (Persson 1975) show that evaporation over the entire winter period amounts only to a few mm. Lemmelä and Kuusisto (1974) found that in southern Finland evaporation from snow did only take place during March and April and was only about 8 mm/winter. Although the evaporation is a negligible part of the water balance, it may be of significance in the energy balance, since the heat needed to evaporate 8 mm of water from snow corresponds to about 60 mm water equivalents of snow melt.

Low evaporation from snow has been confirmed by all careful measurements. The reasons are the infrequency of a strong gradient of vapour pressure from snow to air and/or the low energy input to the snow. Evaporation from snow cover in sub-arctic areas is indeed moderate, since in the winter darkness very little solar energy is available and in spring air currents having the requiste combination of warmth and dryness are infrequent. The extensive snow covers of northern areas loose only small amounts of mass by evaporation, and evaporation from snow has only a minor effect on regional water budgets.

Lysimeter Measurements

At the University of Luleå we have tried to use snow filled pans floating in other larger pans in order to measure the evaporation from snow. The loss of water was directely measured. This method was not proven to be very accurate. The only possible way of measuring the water loss seems to be to measure by weight. The soviet measurements referred to previously are of this type. Also at Luleå this measurement technique has been applied. The pans have, however, only an area of 314 cm^2. The »snow pan« is placed in an outer pan into which melt water is allowed to drain through the perforated bottom of the inner pan. The total evaporimeter or lysimeter is shown in Fig. 6. The weight of the two pans is measured regulary and the amount of melt is also measured. Always four evaporimeters are used at the same time.

During 23 days in the spring of 1979 the average evaporation rate was 0.36 mm/day. Only during 3 days daytime condensation occured. The frequency of evaporation loss during daytime is shown in Fig. 7.

The corresponding measurements were carried out in April 1978 only for 7 days. The daytime, 8-17 hours, evaporation averaged 0.6 mm/day.

Since the evaporation readings were taken at a very well equiped meteorological station it was possible to analyse the measured evaporation in respect to different meteorological parameters. The evaporation was found to be separately correlated to relative humidity, net radiation, dew point temperature, vapour pressure deficit and the standard function $a(1+bw)(e_s-e)$ each with an absolute correlation coefficient, r^2, of 0.8. No separate correlation to the wind speed was found. The measurements fitted the aerodynamic formula best if it was put on the form

$$E = 0.06 + 0.51(1+0.71\,W)(e_s-e) \text{ mm/day} \qquad (24)$$

when W is in m/sec and e is in mb. Since the equation holds only for evaporation during daytime the coefficients 0.06 and 0.51 are consequently reduced to half

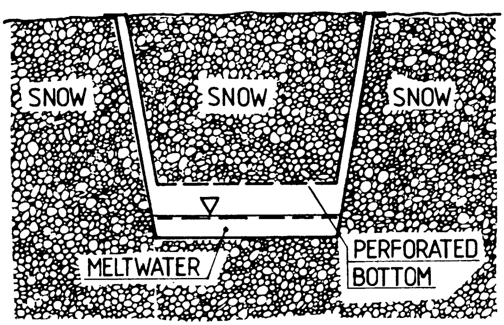

Fig. 6. Snow evaporimeter used at the University of Luleå. Surface area = 0.314 m^2.

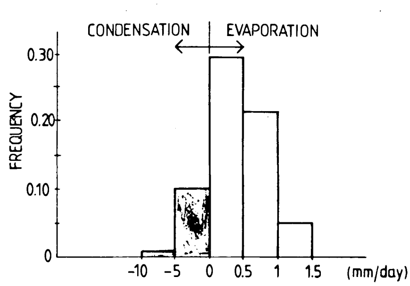

Fig. 7. Frequency of daily evaporation rates from snow evaporimeters during spring 1979 at Luleå.

these values for a period of 12 hours.

It is interesting to note that Lemmelä and Kuusisto (1974) from a similiar regression analysis obtained for 12 hours daytime periods

$$E \equiv 0.03 + 0.11(1+0.86\ W)(e_s - e)\ \text{mm/day} \tag{25}$$

For twelve hours of night time they obtained with a rather poor correlation (r^2 = 0.65).

$$E \equiv -0.02 + 0.17(1+0.06\ W)(e_s - e)\ \text{mm/day} \tag{26}$$

Evaporation as a Transient Process

It was previously pointed out that the heat transfer to/from the snow after a sudden change of air temperature gradually changes from an extreme value towards an equilibrium state. Since energy balance must be obtained the surface temperature must adapt to this balance. For this reason during cold nights with small radiation input the snow surface temperature must drop to low values. Then the air becomes gradually more and more stable. In the morning solar energy becomes available for the snow cover and to obtain energy balance the surface temperature raises up to at the most 0°C and the melting process can start. The fact that high rates of downward heat transfer doesn't warm a melting snow surface means that the gradients of temperature and humidity don't diminish. The air stream above the melting snow is, however, gradually affected so that the temperature immediately above the snow is reduced and consequently also the downward sensible heat flux. The latent heat flux is reduced or increased.

The sensible heat flux from an air stream initially at 5°C to a melting snow surface has purely theoretical been calculated by U. Svensson, University of Luleå, using the momentum equations, the heat energy equation and a k-ε turbulence model. The heat flow as a function of time is shown in Fig. 8 for two different geostrophic winds. The vertical temperature distributions after 6 and 14 hours are shown in Fig. 9.

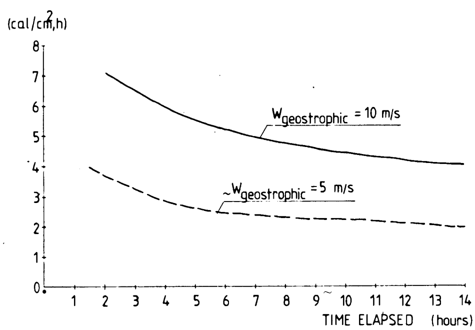

Fig. 8. Sensible heat flux to a melting snow pack from an atmosphere of initial temperature of 5°C.

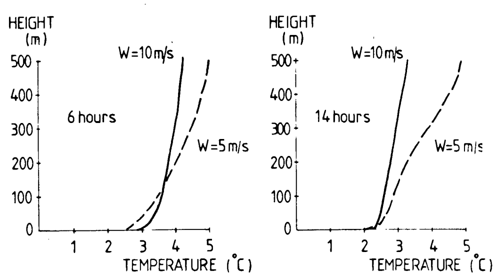

Fig. 9. Vertical air temperature profiles over a melting snow pack after 6 and 14 hours. Initial air temperature +5°C.

Evaporation Related to Upper Atmospheric Conditions

Evaporation is a transient process since radiation, temperature, humidity and wind velocity all affect each other. In order to obtain a satisfactory method of determining evaporation from snow on a routine basis from say some kind of nomogram first a thoroughally analysis should be made. From geostrophic wind, solar radiation, temperature at the base of the cloud cover and humidity the evaporation can be determined theoretically using the well-known conservation laws for heat, mass and momentum. The main difficulty in applying these basic laws stems from the parameterization of turbulence. In recent years it has, however, become possible to predict turbulence exchange coefficients from transport equations of turbulent kinetic energy (k) and dissipation rate (ε) as frequently used by for example Svensson (1978). The temperature profiles previously shown were obtained using such a turbulence model. A model for determining evaporation must, however, include phase exchange in the air and radiation. It is still felt that it should be possible to develop such a model, and that from such a model it would be possible to construct nomograms from which evaporation easily could be determined.

Conclusions

From a water resources point of view evaporation is only of minor importance, but since so much energy is involved in the vaporization process, evaporation highly influences the melting process. For northern areas the total amount of evaporation during the whole snow covered season amounts to 10 or 20 mm. The daily rate of evaporation from snow very seldom exceeds 1 mm. However, evaporation significantly influences the melting rate.

When determining evaporation on a routine basis the aerodynamic formula is most frequently used. It should, however, be used together with energy balance computations, since the evaporation rate is limited by the amount of energy that is available. Also the profile method based on standard coefficients can be used conventionally. This method has the advantage that the effect of density stratification in the air, can be taken into account. However, the method should be used at a rather short time basis.

Although the effect of stratification is taken into account when using the profile method, the change with time of the wind-, temperature - and humidity profiles is neglected. Some known values at for example height 2 m are prerequested. It should be possible to develop a more soundly based model using the conservation laws and a turbulence model so that the evaporation can be described as a transient process.

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