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On the Use of Bulk Aerodynamic Formulae Over Melting Snow

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Bulk aerodynamic formulae which relate the turbulent exchanges of sensible and latent heat over melting snow to measurements of windspeed, temperature and humidity at one level can be derived from flux-gradient relationships and assumed log-linear profiles. Recent analyses of local advection over snow and wind flow over complex terrain suggest that the bulk aerodynamic formulae should apply in non-ideal field situations. The assumption that the scaling lengths for temperature and humidity equal the roughness length is problematic, since theoretical analyses indicate they should be much less than the roughness length. However, the effect of scale length inequality on the stability correction tends to compensate for the effect on the neutral-case transfer coefficient. Field experience indicates that the bulk aerodynamic formulae are adequate for use in energy balance estimates of daily or shorter term snowmelt.

Introduction

Hydrologists are often required to make estimates of snowmelt. The most rigorous method of calculating snowmelt at a point is through the energy balance, in which estimates or measurements are made of the energy exchanges at the snowpack interfaces. Generally, the net radiative exchange provides the greatest portion of the energy disposed of in melt (Male and Granger 1981). However, during weather conditions characterised by strong winds and warm, humid air, the sensible and latent heat fluxes dominate the snowpack energy budget, and produce the highest short-term rates of melt (Hendrick, Filgate and Adams 1971; Light 1941). Thus, accurate and practical methods of evaluating these fluxes are required. The two most accurate methods of determining the turbulent fluxes are the eddy-correlation approach (McKay and Thurtell 1978) and detailed profile methods (de la Casiniere 1974). Unfortunately, both these methods require extensive and expensive instrumentation, and are not well suited to the conditions and environments often of interest to hydrologists. In addition, neither method is readily suitable for telemetry applications for real-time streamflow forecasting. In practice, therefore, hydrologists often resort to methods which require measurements of wind speed, temperature and humidity at one level and exploit the fact that the temperature and vapour pressure at a melting snow surface are 0°C and 6.11 mb.

One method for calculating the sensible and latent heat fluxes uses empirical wind functions (e.g., Anderson 1976). The wind function depends on stability, surface roughness and measurement height. The empirically-derived wind functions will represent the average surface and stability conditions prevailing during the calibration period, and will be biased. Anderson (1976) and Male and Gray (1981) attempted to compare coefficients derived by several workers by reducing the measurement levels to a standard height of 1 m. The choice of extrapolation formula is arbitrary, so comparisons cannot be conclusive. However, the comparisons indicate that the coefficients range greatly between environments, as would be expected.

Another method of estimating the turbulent exchanges over melting snow is the bulk aerodynamic method, which employs bulk transfer coefficients derived from flux-gradient relationships and semi-empirical profile forms for wind, temperature and humidity. These bulk transfer coefficients explicitly account for variable stability, surface conditions and measurement heights, and have been used by Anderson (1976), Price (1977), Heron and Woo (1978), Prowse and Owens (1982) and Moore and Owens (in press). However, several important assumptions underlying the bulk aerodynamic approach are not discussed adequately by any of these authors. This paper presents a derivation of the bulk aerodynamic transfer coefficients, discusses some of the assumptions and their practical implications, and reviews the field evidence for the validity of the bulk aerodynamic approach.

Theory

The exchanges of momentum (M), sensible heat (Q_H) and latent heat (Q_E) are often expressed in the surface layer by the familiar flux-gradient equations

$$M \equiv \rho K_M \frac{du}{dz} \tag{1a}$$

$$Q_H = \rho c_p K_H \frac{dT}{dz} \tag{1b}$$

$$Q_E = \rho L_V K_E \frac{dq}{dz} \tag{1c}$$

where

M is in N m^{-2}, and Q_H and Q_E are in W m^{-2}
ϱ is the air density in kg m^{-3}
K_M, K_H and K_E are the eddy diffusivities for momentum, sensible heat and water vapour in m^2 s^{-1}
c_p is the specific heat of air at constant pressure in J kg^{-1} °C^{-1}
L_V is the latent heat of vaporisation in J kg^{-1}
T is air temperature in °C
q is the specific humidity in g g^{-1}
z is the height above the surface in m

Contrary to micrometeorological convention, the fluxes are here considered positive when directed toward the surface. Eqs. (1) can be expressed in integrated form as (after Deardorff 1968)

$$M \equiv \rho D_M U_\alpha \tag{2a}$$

$$Q_H = \rho c_p D_H (T_\alpha - T_s) \tag{2b}$$

$$Q_E = \rho L_V D_E (q_\alpha - q_s) \tag{2c}$$

where D_M, D_H and D_E are the bulk exchange coefficients for momentum, heat and water vapour in m s^{-1} the subscripts s and a refer to the values at the surface and at height z_α.

In practice, the term $(q_\alpha - q_s)$ is replaced by $(0.622/p)$ $(e_\alpha - e_s)$, where p is atmospheric pressure (mb) and e is the vapour pressure (mb).

The snow surface temperature has an upper limit of 0°C. During snowmelt, the air temperature is usually greater than 0°C, so atmospheric conditions over melting snow are generally stable (de la Casiniere 1974). Log-linear profiles for wind, temperature and humidity apply in this stability range (Webb 1970), and the gradients can be written

$$\frac{du}{dz} = \frac{u^*}{kz} \phi_M\left(\frac{z}{L}\right) \tag{3a}$$

$$\frac{dT}{dz} = \frac{T^*}{kz} \phi_H\left(\frac{z}{L}\right) \tag{3b}$$

$$\frac{dq}{dz} = \frac{q^*}{kz} \phi_E\left(\frac{z}{L}\right) \tag{3c}$$

$$\phi_i\left(\frac{z}{L}\right) = 1 + b_i \frac{z}{L}$$

where

u^* is the friction velocity $= (M/\varrho)^{1/2}$ in m s^{-1}
T^* and q^* are temperature and humidity scales analogous to friction velocity

k is von Karman's constant, equal to 0.4
b_i are empirical constants
L is the Monin-Obukov length scale in m.

The length scale, L, is defined by

$$L = \frac{T_k u^{*3} \rho c_p}{k g Q_H}$$

where

T_K is the mean temperature of the air layer in K
g is the gravitational acceleration in m s^{-2}

The eddy diffusivities can be assumed equal for near-neutral to stable conditions (Oke 1970). Equality of diffusivities implies that $\phi_M = \phi_H = \phi_E$; that is, $b_M = b_H = b_E = b$ (Dyer 1974). Dyer (1974) noted that available evidence suggests that $b = 5$. If the eddy diffusitivies are equal, Eqs. (3) can be integrated to yield

$$u_\alpha = \frac{u^*}{k}(\ln\frac{z_\alpha}{z_0} + b\frac{z_\alpha}{L}) \qquad (4a)$$

$$T_\alpha - T_s = \frac{T^*}{k}(\ln\frac{z_\alpha}{z_T} + b\frac{z_\alpha}{L}) \qquad (4b)$$

$$q_\alpha - q_s = \frac{q^*}{k}(\ln\frac{z_\alpha}{z_q} + b\frac{z_\alpha}{L}) \qquad (4c)$$

where

$z_\alpha >> z_o, z_T, z_q$
z_o is the roughness length in m
z_T and z_q are scaling heights for T and q, analogous to z_o, in m.

and the quantities u^*, T^* and q^* are defined by the following expressions (Busch 1973)

$$M = \rho u^{*2} \qquad (5a)$$

$$Q_H = \rho c_p u^* T^* \qquad (5b)$$

$$Q_E = \rho L_V u^* q^* \qquad (5c)$$

Male and Granger (1981) noted that consensus amongst micrometeorologists is divided as to the profile forms for temperature and humidity. Many investigators support an alternative formulation due to Businger, Wyngaard, Izumi and Bradley (1971), in which the right hand sides of Eqs. (4b) and (4c) are multiplied by a factor $R = 0.74$, b is replaced by b/R and $k = 0.35$. However, a re-analysis of the data set used by Businger *et al.* and an analysis of a new data set support the more traditional formulation given by Eqs. (4b) and (4c), with $k = 0.4$ (Wieringa 1980; Dyer and Bradley 1982).

The bulk exchange coefficients are derived by combining Eqs. (2), (4) and (5); that is

$$D_M = \frac{k^2 u_\alpha}{(\ln \frac{z_\alpha}{z_0} + b \frac{z_\alpha}{L})^2} \tag{6a}$$

$$D_H = \frac{k^2 u_\alpha}{(\ln \frac{z_\alpha}{z_0} + b \frac{z_\alpha}{L})(\ln \frac{z_\alpha}{z_T} + b \frac{z_\alpha}{L})} \tag{6b}$$

$$D_E = \frac{k^2 u_\alpha}{(\ln \frac{z_\alpha}{z_0} + b \frac{z_\alpha}{L})(\ln \frac{z_\alpha}{z_q} + b \frac{z_\alpha}{L})} \tag{6c}$$

Before Eqs. (6) can be applied, the scale heights z_o, z_T and z_q and the stability parameter z_α/L must be known; or alternatively, if z_o is known and T and q are measured at two levels, the fluxes can be calculated iteratively, initially assuming $z_\alpha/L = 0$ (Berkowicz and Prahm 1982). For the case of measurements at only one level, another approach is to assume equality of the scale heights and relate z_α/L to a more readily computed stability parameter.

A commonly used stability parameter is the gradient Richardson number, defined by

$$Ri \equiv \frac{g \frac{dT}{dz}}{T_K (\frac{du}{dz})^2} \tag{7}$$

If $z_T = z_o$ (or equivalently, if $T_o = T(z_o) = T_s$), the bulk form of Ri can be defined as

$$Rb = \frac{g(T_\alpha - T_s)(z_\alpha - z_0)}{T_K u_\alpha^2} \tag{8}$$

If $K_H = K_M$, then Oke (1970) shows that

$$\frac{z}{L} = \frac{Ri}{1 - b\,Ri} \tag{9}$$

From Eqs. (3), (4), (7) and (8), we derive

$$Ri = \frac{Rb(\ln\frac{z_\alpha}{z_0} + b\frac{z_\alpha}{L})}{1+b\frac{z_\alpha}{L}} \quad (10)$$

Combining Eqs. (9) and (10),

$$\frac{z_\alpha}{L} = \frac{Rb\ \ln\frac{z_\alpha}{z_0}}{1 - b\ Rb} \quad (11)$$

If $z_T = z_q = z_o$, then $D_H = D_E = D_M$. Under neutral conditions $z/L = 0$, so the ratio of the stable and neutral case transfer coefficients (D_S/D_N) is

$$\frac{D_S}{D_N} = \frac{(\ln\frac{z_\alpha}{z_0})^2}{(\ln\frac{z_\alpha}{z_0} + b\ \frac{z_\alpha}{L})^2} \quad (12)$$

Substituting Eq. (11) into Eq. (12) gives

$$\frac{D_S}{D_N} = (1 - b\ Rb)^2 \quad (13)$$

Therefore, the bulk exchange coefficients are given by

$$D_H = D_E = D_M = \frac{k^2 u_\alpha (1-b\ Rb)^2}{(\ln\frac{z_\alpha}{z_0})^2} \quad (14)$$

Deardorff (1968) derived ratios of the bulk transfer coefficients for the general case of $D_H \neq D_E \neq D_M$. Anderson (1976) noted that Deardorff's ratios are the same as Eq. (13) if the bulk transfer coefficients are equal. Price (1977) derived ratios of $1/(1 + 10\ Rb)$ for stable conditions and $(1 - 10\ Rb)$ for unstable conditions. These ratios are equivalent to Eq. (13) for small values of Rb, since his derivation assumed that z_α/L is small.

The only unknown quantity in Eq. (14) is z_o. The value of z_o can be calculated for neutral conditions from wind profiles or from wind measurements at two heights z_1 and z_2 according to

$$z_0 = \exp(\frac{u_2 \ln z_1 - u_1 \ln z_2}{u_2 - u_1}) \quad (15)$$

This formula is sensitive to errors in wind speed, so the scatter is usually large, and

Table 1 – Roughness lengths over melting snow

z_o (mm)	Remarks	Source
2.5	glacier snow, 1	Sverdrup (1936)
0.9	glacier snow, 1	Wendler and Streten (1969)
0.9	glacier snow, 1	Wendler and Weller (1974)
5.0	seasonal snow, 2	Price (1977)
5.0	suncupped glacier snow, 2	Fohn (1973)
4.0	seasonal snow, 3	Moore and Owens (in press)
5.0	seasonal snow, 4	Anderson (1976)
4.0	seasonal snow, 1	Grainger and Lister (1966)
6.8	"undulating wet snow", 1	Grainger and Lister (1966)
3.0	seasonal snow, 3	Heron and Woo (1978)
.2-.8	level snow cover, 1	Konstantinov (1966)
1-5	average snow cover, 1	Konstantinov (1966)
5-20	weak friable snow, 1	Konstantinov (1966)

Notes:
1. Calculated from profiles.
2. Calculated from Lettau's (1969) formula.
3. Calculated from wind speeds at two heights.
4. Based on calibration of snowmelt model (with stability correction) with lysimeter outflow.

an average value from several runs is required (Berkowicz and Prahm 1982). Alternatively, z_o has been calculated from Lettau's (1969) formula based on a description of the surface roughness elements (Fohn 1973; Price 1977). Table 1 shows values of z_o over melting snow given in the literature. The values mostly lie in the range from 1 mm, corresponding to "smooth" snow, to 5 mm, for drifted or suncupped snow. Fig. 1 illustrates the dependece of the bulk transfer coefficients on z_o. The figure shows that if the roughness length can be estimated to within 1 or 2 mm, the resulting error in the calculated flux will be within about 20 %.

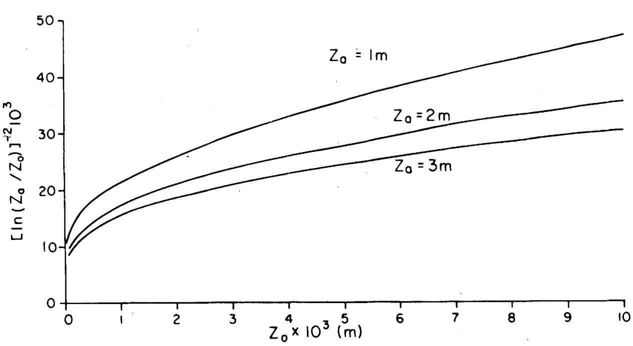

Fig. 1. The dependence of the bulk transfer coefficient on the roughness length.

Assumptions

Three assumptions, besides the validity of the flux-profile relationships, were made in the derivation of Eq. (14). These are the validity of log-linear profiles, the equality of the eddy diffusivities, and the equality of the scale heights for windspeed, temperature and humidity. These assumptions and their implications will be discussed in turn.

Log-Linear Profiles

The derivation of log-linear profiles assumes that there is no radiative or turbulent flux divergence in the atmospheric surface layer (Lumley and Panofsky 1964). During weather situations characterised by light winds, the profiles of wind, temperature and humidity deviate markedly from log-linear forms due to radiative flux divergence and the lack of vertical mixing (Halberstam and Schieldge 1981). However, under these conditions, the sensible and latent heat fluxes are less important than radiation, so inaccuracies in their estimation should not be critical to snowmelt computations. The constant turbulent flux assumption also requires a homogeneous surface and steady state conditions (Lumley and Panofsky 1964). The satisfaction of the steady state requirement is difficult. Brutsaert (1982) suggested the use of averaging times from 20 to 60 min as a compromise between achieving sampling stability and ensuring stationarity of atmospheric conditions. Joffre (1982) discussed stationarity problems in more detail. Homogeneous conditions are rarely met in the field. Fortunately, snowcover tends to smooth small-scale terrain inhomogeneity and produce a homogeneously rough surface. Despite this smoothing by snowcover, two types of inhomogeneity can still exist.

One common form of surface inhomogeneity is a leading edge between snow-free and snow-covered ground. In this situation, horizontal flux divergence occurs, violating one of the assumptions of the aerodynamic approach. However, a numerical analysis by Weisman (1977) indicates that most of the flux divergence occurs within several metres downwind of the leading edge, so that one-dimensional equations such as Eq. (2) should be applicable at the measurement site given a sufficient fetch. Male and Gray (1981) suggested that the height of the constant flux layer is one-hundredth the fetch. Thus, for instrument heights of 1 to 2 m, an upwind fetch of 100 to 200 m should be sufficient.

Terrain features such as slopes and hills affect the structure of turbulence. The analysis by Jackson and Hunt (1975) shows that these effects can be neglected for flow over hills with slopes on the order of 0.05. Furthermore, if the velocity profile upwind of the hill is logarithmic, then it remains logarithmic with the same roughness length in the inner layer as the flow passes over the hill, even though the airflow accelerates (Jensen and Peterson 1978). This statement does not apply to the separated-flow zone on the downwind side of the hill. Field observations by

Mason and Sykes (1979) suggest that the results of Jackson and Hunt apply to hills with slopes much greater than assumed by their theory. There is as yet no consensus as to the effect of flow over hills on a log-linear upwind profile (A. J. Bowen pers. comm.). These results suggest that the aerodynamic approach should be applicable even in the non-ideal field situations confronting most researchers, as long as the terrain is not too complex and the fetch is sufficient.

Equality of Diffusivities

The equality $K_H = K_E = K_M$ is supported in the stable range by several studies (Dyer 1974), but has been questioned by some workers (Male and Gray 1981). The equality of K_H and K_M is supported by most studies,at least for the range of stability in which turbulent exchange is important. However, as mentioned previously, many researchers support the profile formulations of Businger *et al.* (1971), which imply that $K_H/K_M = 1.35$ under neutral conditions, and that von Karman's constant equals 0.35. Unfortunately, the evidence supporting either set of formulations is still inconclusive. The equality of K_E and K_H in the stable range is supported by studies such as Oke (1970) and Webb (1970), while Male and Granger (1981) reported results that show $K_E/K_H = 0.5$ for most of the stable range. Lang, McNaughton, Fazu, Bradley and Ohtaki (1983) presented evidence that $K_E > K_H$ when Q_H is towards the surface and Q_E is away from the surface. However, during many rapid melt situations, Q_H and Q_E are both directed toward the surface, and there appear to be no detailed studies of this situation. Male and Granger (1981) suggest that this controversy might be explained by McBean and Miyake's (1972) statement that universal transfer laws for passive scalars such as moisture may not exist. The evidence against equality of the eddy diffusitivies in the stable range is inconclusive, so the assumption that $K_H = K_E = K_M$ is as reasonable as any other, but carries some degree of uncertainty.

Equality of Scale Heights

The equality of the scale heights is the most problematic assumption, despite the arguments by Joffre (1982) that $z_T = z_o$. Very few values for z_T over melting snow have been given in the literature. Sverdrup (1936) reported values from .072 to .32 mm (z_o = 2.5 mm), while Wendler and Streten (1966) reported $z_T = z_o = .9$ mm. No values for z_q over melting snow were found, probably because of the difficulty of obtaining accurate measurements of humidity. Unlike z_o, which is primarily a function of surface characteristics, z_T and z_q depend upon the thickness of the laminar boundary layer, and vary with flow conditions. Webb (1975) stated that z_T and z_q equal $D/(nku^*)$, where D is the molecular diffusivity in air (m^2s^{-1}) and n is a parameter equal to 7 for water vapour. Substitution of typical values of u^* over melting snow into this expression yields scale heights one to two orders of magnitude smaller than z_o. Values for z_T and z_q calculated from an alternative expression due to Brutsaert (1975) are similar. Fig. 2 illustrates the dependence of the neu-

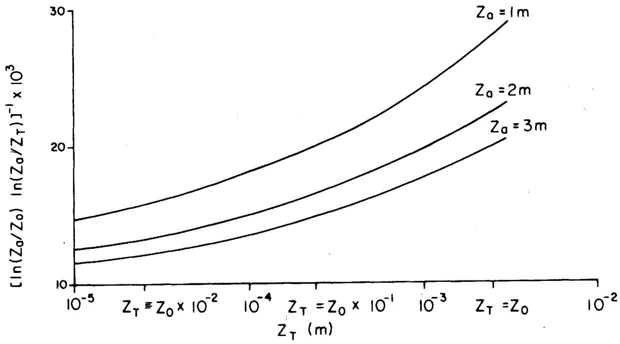

Fig. 2.
The dependence of the neutral case bulk transfer coefficient on the temperature scaling length for $z_o \equiv .0025$.

tral-case bulk transfer coefficient on a variable z_T, given $z_o = 2.5$ mm. If $z_T = z_o/10$, assuming that $z_T = z_o$ will produce an overestimate of 30-40 %, depending on the measurement height; and if $z_T = z_o/100$, the overestimate will be 65-75 %. However, as discussed in the next section, the studies which have employed bulk aerodynamic formulae do not report a positive bias in the snowmelt computed from the energy budget, which would be expected from the considerations given above.

At least a partial answer to the problem posed by unequal scale heights can be found by examining their effect on the stability correction. Consider a situation in which the profiles of windspeed, temperature and humidity are given by Eqs. (4). The value of *Rb* which is calculated from the measurements and is used to compute the stability correction is given by

$$Rb_C = \frac{Ri(1+b\frac{z}{L})(\ln\frac{z}{z_T} + b\frac{z}{L})}{(\ln\frac{z}{z_0} + b\frac{z}{L})^2},$$

where *Ri* and *z*/*L* are given for any height from the profiles. The value of *Rb* assumed by the derivation of the stability correction is given by

$$Rb_A = \frac{Ri(1+b\frac{z}{L})}{\ln\frac{z}{z_0} + b\frac{z}{L}}$$

The ratio of Rb_C to Rb_A is

$$\frac{Rb_C}{Rb_A} = \frac{\ln\frac{z}{z_T} + b\frac{z}{L}}{\ln\frac{z}{z_0} + b\frac{z}{L}} \qquad (16)$$

Eq. (11) was solved to calculate Rb_A as a function of z/L; Rb_C was then calculated from Eq. (16). Fig. 3 shows the ratios of the stability correction calculated from Rb_C to that calculated from Rb_A as a function of Rb_C. As Rb_C and stability

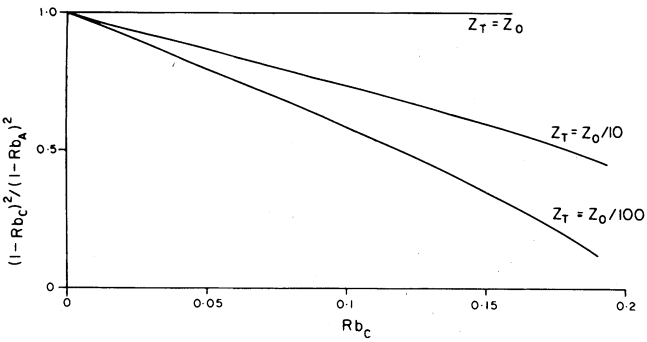

Fig. 3.
Ratio of the stability corrections using calculated and assumed values of Rb.

increase, the ratio of the corrections decreases, tending to compensate for the overestimation caused by assuming $z_T = z_q = z_0$. The point at which the bulk transfer coefficient overestimation and the correction factor underestimation compensate for each other is at $Rb_C = .095$ for $z_T = z_o/10$ and at $Rb_C = .098$ for $z_T = z_o/100$. Anderson (1976) found that turbulent exchange was most significant as an energy source for snowmelt for Rb_C in the range .05 to .09. In this range, the positive bias due to variations from $z_T = z_0$ would average on the order of 10 %, and could be masked by instrument error or uncertainty in the value of z_0. However, more research is required on the behaviour of z_T and z_q over melting snow, and the implications for heat and vapour transfer.

Field Experience

Several studies have employed bulk aerodynamic formulae in conjunction with measurements or estimates of the snow surface radiative exchange to compute snowmelt. Anderson (1976) compared computed snowmelt with lysimeter outflow for both daily and hourly periods. The agreement in both cases was good, with a negative bias of 6 %. Heron and Woo (1978) found on a daily basis that melt estimated from the energy budget agreed to within 25 % of melt computed from depth and density measurements, except during the initial melt period, when the snowpack was ripening, and during the late melt period, when the snowcover was patchy. Moore and Owens (in press) found that computed snowmelt over an eight day period agreed to within 3 % of that calculated from depth and density measurements. Price (1977) found good agreement between computed melt and runoff from a hillslope on a daily basis, with a standard error about the regression line (which was not significantly different from the 1:1 line) of 7 mm/d.

There are two problems associated with using these comparisons to verify the bulk aerodynamic formulae. First, errors in the computed melt may be caused by errors in the radiative exchange, which in most of the cases was the greatest source of energy for snowmelt. Second, the errors involved in determining daily ablation from depth and density are possibly of the same magnitude as the errors in daily computed snowmelt.

Male and Granger (1979) compared daily values of Q_H and Q_E computed from bulk aerodynamic formulae with "measured" values. Their comparison indicates that the bulk aerodynamic formulae are not accurate, and overestimate evaporation. They attributed the scatter to the radiative heating of the lower air layers, which would cause profiles of windspeed, temperature and humidity to deviate from log-linear forms. It is difficult to assess this comparison because Male and Granger did not give details of their computation, such as the value of z_o used and the averaging time for the values of windspeed, temperature and humidity; nor did they state how they measured Q_H and Q_E. In conditions similar to those for which this comparison was made, Q_H and Q_E are usually much less important than radiative exchange for snowmelt. The conditions in which Q_H and Q_E are important to the energy budget are often characterised by moderate to strong winds and low insolation – conditions in which log-linear profiles are most likely to apply.

Conclusions

Formulae relating the turbulent exchanges of sensible and latent heat over melting snow to measurements of windspeed, temperature and humidity at one level can be derived from flux-gradient relationships and assumed profile forms. These bulk aerodynamic formulae are superior to empirical wind functions because they can account for differences in stability, surface conditions and measurement heights, and do not have to be empirically calibrated. The main problem in their application is in specifying the roughness length. Fortunately, the roughness length ranges over less than an order of magnitude, and the formulae are not critically sensitive to changes in the roughness length. Field experience indicates that the bulk aerodynamic formulae are adequate for use in energy budget computations of daily or shorter-term snowmelt. However, bulk aerodynamic computations of Q_H and Q_E need to be tested directly against more accurate methods under a variety of conditions to make clear their limitations.

The most problematic assumption underlying the bulk aerodynamic approach is the equality of the scale heights for wind, temperature and humidity. More research needs to be carried out on the behaviour of the scale heights for temperature and humidity over melting snow, and the implications for heat and vapour transfer.

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