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On the exchange of sensible and latent heat between the atmosphere and melting snow

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ABSTRACT

The snow energy balance is difficult to measure during the snowmelt period, yet critical for predictions of water yield in regions characterized by snow cover. Robust simplifications of the snowmelt energy balance can aid our understanding of water resources in a changing climate. Research to date has demonstrated that the net turbulent flux (F_T) between a melting snowpack and the atmosphere is negligible if the sum of atmospheric vapor pressure (e_a) and temperature (T_a) equals a constant, but it is unclear how frequently this situation holds across different sites. Here, we quantified the contribution of F_T to the snowpack energy balance during 59 snowmelt periods across 11 sites in the FLUXNET2015 database with a detailed analysis of snowmelt in subarctic tundra near Abisko, Sweden. At the Abisko site we investigated the frequency of occurrences during which sensible heat flux (H) and latent heat flux (λE) are of (approximately) equal but opposite sign, and if the sum of these terms, F_T, is therefore negligible during the snowmelt period. H approximately equaled -λE for less than 50% of the melt period and F_T was infrequently a trivial term in the snowmelt energy balance at Abisko. The reason is that the relationship between observed e_a and T_a is roughly orthogonal to the "line of equality" at which H equals -λE as warmer T_a during the melt period usually resulted in greater e_a. This relationship holds both within melt periods at individual sites and across different sites in the FLUXNET2015 database, where F_T comprised less than 20% of the energy available to melt snow, Q_m, in 44% of the snowmelt periods studied here. F_T/Q_m was significantly related to the mean e_a during the melt period, but not mean T_a, and F_T tended to be near 0 W m^{-2} when e_a averaged *ca.* 0.5 kPa. F_T may become an increasingly important term in the snowmelt energy balance across many global regions as warmer temperatures are projected to cause snow to melt more slowly and earlier in the year under conditions of lower net radiation (R_n). Eddy covariance research networks such as *Ameriflux* must improve their ability to observe cold-season processes to enhance our understanding of water resources and surface-atmosphere exchange in a changing climate.

1. Introduction

Quantifying the snowpack energy balance is critical for water regulation and runoff prediction (Dettinger et al., 2015; Kay and Crooks, 2014; Marks et al., 2008; Troin et al., 2016), avalanche forecasting (Slaughter et al., 2009; Wever et al., 2016), and predicting changes to future snowpack persistence (Abatzoglou et al., 2014; Pederson et al., 2011). Interannual variability in weather and climate change impact the timing and magnitude of snowmelt (Cline, 1997; Grundstein and Leathers, 1999; Hayashi et al., 2005; Pederson et al., 2013), and snowmelt is projected to occur earlier and more slowly in a warming climate under conditions of lower net radiation earlier in the season (Musselman et al., 2017). To understand how snowmelt responds to climate variability, we must understand mass and energy fluxes to and from the snowpack, including key relationships that can simplify models without impacting their skill.

Using the convention that energy flux into the snowpack is positive, the energy available to melt snow, Q_m, is a function of the net radiation

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(R_n, i.e. incident minus outgoing shortwave and longwave radiation), sensible heat flux (H), latent heat flux (λE), ground heat flux (G), and any energy flux due to precipitation (P, Marks and Dozier, 1992; Burns et al. 2014):

$$Q_m + Q_{cc} = R_n + H + \lambda E + G + P = R_n + F_T + G + P. \quad (1)$$

Q_{cc}, the energy required to bring snow temperature to melting temperature (often called the cold content), is assumed here to be 0 W m^{-2} when the snowpack is melting. The net turbulent flux, F_T, is the sum of H and λE, the latter being of particular interest to snow science as it represents sublimation and evaporation from and condensation to the snowpack, and is thus connected to the snow mass balance. H and λE tend to be minor but nontrivial contributions to Q_m (Boon, 2009; Cline, 1997; Harding and Pomeroy, 1996; Fitzpatrick et al. 2017; Marks and Winstral, 2001) and are arguably more difficult to measure via, for example, eddy covariance compared to the radiometers and heat flux plates used to measure R_n and G (Arck and Scherer, 2002). Understanding situations in which the contribution of F_T to Q_m is negligible would dramatically simplify our ability to measure and model Q_m.

In a comparison of studies at alpine sites over portions of the melt period, Cline (1997) found that the contribution of R_n to Q_m ranged from 0 to 100%, illustrating that F_T can be both a negligible and dominant source of energy for snowmelt. F_T can dominate Q_m in arid environments, especially in the early season before R_n reaches higher values (Beaty, 1975; Hawkins and Ellis, 2007). The contribution of F_T to Q_m can vary due to weather patterns (Cline, 1997; Grundstein and Leathers, 1999; Hayashi et al., 2005), wind speed (Mott et al., 2011; Pohl et al., 2006), and vegetation (Endrizzi and Marsh, 2010; Mahrt and Vickers, 2005), which makes model simplification difficult. However, the time scales of these comparisons range from the entire snow covered period to less than a week, and it is unclear how frequently, and under which conditions, F_T contributes negligibly to Q_m when snow is melting.

Welch et al. (2016) used eddy covariance measurements in a montane continental snowpack in Montana, USA, and found that H was only about 10% less than the magnitude of -λE, such that F_T (-3 MJ m^{-2}) provided a negligible contribution to Q_m (97 MJ m^{-2}) when integrated over the entire melt period. They described a linear relationship between near-surface air temperature (T_a) and atmospheric vapor pressure (e_a) for conditions under which $H = -\lambda E$ (Fig. 1, i.e. $F_T = 0$ W m^{-2} and the Bowen ratio $\beta = H/\lambda E = -1$) that results when snow surface temperature (T_{ss}) is at 0 °C when snow is melting. It was noted that average T_a and e_a during the melt period fell near the line at

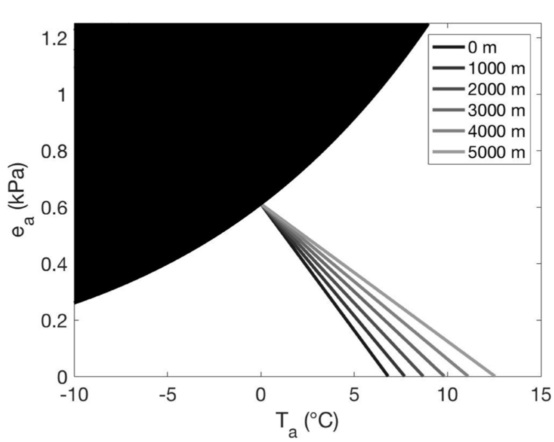

Fig. 1. The atmospheric vapor pressure (e_a) and temperature (T_a) at which energy flux into snow from sensible heat (H) and losses from latent heat (λE) during snowmelt are equal (i.e. $H = -\lambda E$) following Welch et al. (2016) for different elevations above sea level and therefore mean psychrometric constants (γ), which are a function of atmospheric pressure with minor temperature dependency as discussed in Loescher et al. (2009). The black shaded area denotes the region for which e_a exceeds saturation over ice following Goff and Gratsch (1945).

which $H = -\lambda E$, hereafter the "line of equality", and derived below in *Methods*. It is unclear if other melting snowpacks experience similar average climate conditions that make F_T negligible during the melt period and therefore when R_n measurements alone provide an accurate approximation of Q_m (Eq. (1), noting that the magnitude of G is often trivial compared to other terms in Eq. (1) during snowmelt.

Here, we quantify the contribution of F_T to Q_m during two snowmelt periods at a subarctic tundra research site near Abisko, Sweden and 59 snowmelt periods across 11 eddy covariance study sites in the FLUXNET2015 database (Pastorello et al., 2017) to quantify the range of meteorological conditions encountered during the melt period in different snowpacks. We examined two questions. First, are the micrometeorological conditions during the snowmelt period observed in Welch et al. (2016) common for various sites with different physical characteristics? To address this question, we examined eddy covariance and radiometric measurements of energy exchange between the snowpack and the atmosphere from sites in different climate zones. Second, how frequently is F_T approximately equal to 0 W m^{-2}, and what is the relative contribution of F_T to Q_m during the snowmelt period across sites? To address this question we study the distribution of micrometeorological conditions during different melt events. The goal of this analysis is to gain a better understanding of conditions in which the snowmelt energy balance can be accurately approximated using radiometric observations to simplify measurements and models. We focus our discussion on the steps necessary to improve observations of cold season processes within surface-atmosphere flux networks like *Ameriflux,* and to improve observations of climate and surface-atmosphere flux at snowmelt measurement networks like SNOTEL (snowpack telemetry, Serreze et al., 1999).

2. Methods

2.1. Snow energy balance and turbulent flux during snowmelt

Welch et al. (2016) present a relationship in which the input of H to the snowpack is equal to $-\lambda E$ (i.e. $F_T = 0$ W m^{-2}) when snow is melting. Briefly, H can be written following e.g. Kaimal and Finnigan (1994):

$$H = \frac{\rho C_p}{r_H}(T_a - T_{ss}) \quad (2)$$

where ρ is the molar density of air measured in mol m^{-3}, C_p is the specific heat of dry air (J mol^{-1} K^{-1}), and r_H is the resistance to heat flux (s m^{-1}). H is positive when T_a exceeds T_{ss} noting the convention here that energy flux from the atmosphere to the snowpack is positive; positive H denotes heat transport from the surface to the atmosphere in conventional flux studies. Welch et al. (2016) assumed that T_{ss} is 0 °C (273.15 K) when snow was melting. Thus,

$$H = \frac{\rho C_p T_a}{r_H} \quad (3)$$

We note that T_{ss} in Eq. (2) is the *aerodynamic* surface temperature, which is the temperature that influences turbulent flow. The aerodynamic surface temperature is similar to radiative surface temperature if melting snow can be considered a smooth surface for the case of eddy covariance research sites, and is thus related to the outgoing longwave radiation (LW_{out}) following the Stefan-Boltzmann equation:

$$LW_{out} = A\varepsilon\sigma T_{sr}^4 \quad (4)$$

where A is the view factor (assumed to be 1 for a melting snowpack on a flat surface), σ is the Stefan-Boltzmann constant, ε is the emissivity of snow (which changes as a function of snow characteristics, e.g. Hori et al., 2006), and T_{sr} is the radiometric snow surface temperature in degrees Kelvin, 273.15 K if snow is melting. Aerodynamic and radiometric surface temperatures are otherwise different terms.

Like H, λE can be written: