



Comparing Simulated and Measured Sensible and Latent Heat Fluxes over Snow under a Pine Canopy to Improve an Energy Balance Snowmelt Model

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ABSTRACT

During the second year of the NASA Cold Land Processes Experiment (CLPX), an eddy covariance (EC) system was deployed at the Local Scale Observation Site (LSOS) from mid-February to June 2003. The EC system was located beneath a uniform pine canopy, where the trees are regularly spaced and are of similar age and height. In an effort to evaluate the turbulent flux calculations of an energy balance snowmelt model (SNOBAL), modeled and EC-measured sensible and latent heat fluxes between the snow cover and the atmosphere during this period are presented and compared. Turbulent fluxes comprise a large component of the snow cover energy balance in the premelt and ripening period (March–early May) and therefore control the internal energy content of the snow cover as melt accelerates in late spring. Simulated snow cover depth closely matched measured values (RMS difference 8.3 cm; Nash–Sutcliffe model efficiency 0.90), whereas simulated snow cover mass closely matched the few measured values taken during the season. Over the 927-h comparison period using the default model configuration, simulated sensible heat H was within 1 W m^{-2} , latent heat $L_v E$ within 4 W m^{-2} , and cumulative sublimation within 3 mm of that measured by the EC system. Differences between EC-measured and simulated fluxes occurred primarily at night. The reduction of the surface layer specification in the model from 25 to 10 cm reduced flux differences between EC-measured and modeled fluxes to 0 W m^{-2} for H , 2 W m^{-2} for $L_v E$, and 1 mm for sublimation. When only daytime fluxes were compared, differences were further reduced to 1 W m^{-2} for $L_v E$ and $<1 \text{ mm}$ for sublimation. This experiment shows that in addition to traditional mass balance methods, EC-measured fluxes can be used to diagnose the performance of a snow cover energy balance model. It also demonstrates the use of eddy covariance methods for measuring heat and mass fluxes from snow covers at a low-wind, below-canopy site.

1. Introduction

Water from melting snow is a critical resource in western North America and other similar regions of the world. Across the intermountain western United States, most of the landscape is arid or semiarid, receiving less than 30 cm of annual precipitation. About 15% of the land area is above 2000 m, and it generally receives substantially more annual precipitation, 70%–90% of which has historically fallen as snow (Anderson et al. 1976). The seasonal snow cover has acted as a natural

reservoir to store water from winter storms for spring and summer delivery to soils and streams in the region. In recent years, however, notable changes in the seasonal climate and snow cover have been observed (Mote et al. 2005; Hamlet et al. 2005), characterized by warmer temperatures, reduced winter snowfall, increased winter rainfall, earlier peak snow accumulation and melt, and, in many areas, reduced late spring and summer precipitation. These changes will put additional stress on already limited water resources in the western United States (Barnett et al. 2005) and will require improved monitoring (Schaefer and Werner 1996; Abramovich and Pattee 1999). Furthermore, as empirical methods calibrated on past climate conditions become less reliable, a more physically based spatially explicit approach to forecasting melt from the seasonal

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snow cover across the region (Garen and Marks 1998, 2005) is essential.

A number of studies have focused on both measuring and modeling the snow cover energy and mass balance (e.g., Anderson 1976; Male and Granger 1981; Marks et al. 1992; Marks and Dozier 1992; Harding and Pomeroy 1996; Marsh and Pomeroy 1996; Pomeroy et al. 1998; Hedstrom and Pomeroy 1998; Marks et al. 1998; Luce et al. 1998, 1999; Tarboton et al. 2000; Marks and Winstral 2001; Tribbeck et al. 2004; Pomeroy et al. 2002, 2003; Essery et al. 1998). All these studies show that radiation and turbulent fluxes dominate the snow cover energy balance and all indicate that modeled turbulent fluxes are difficult to validate. Below-canopy radiation fields over snow have been measured and validated (e.g., Pomeroy and Dion 1996; Melloh et al. 2001; Hardy et al. 2004; Link et al. 2004; Sicart et al. 2004; Tribbeck et al. 2006; Pomeroy et al. 2008). However, only with the recent development of robust, low-power eddy covariance (EC) instrumentation has direct measurement of turbulent fluxes between the snow cover and the atmosphere been possible for periods of time long enough to validate snow cover energy and mass balance. Harding and Pomeroy (1996) showed that forest canopies substantially alter turbulent fluxes and that energy closure is difficult with mixed surfaces of snow, evergreen canopies, and trunks. Pomeroy et al. (1998) show that substantial errors can accrue to turbulent flux estimates from models in which the surface temperature and internal energetics of the model are in error.

Over a snow cover, few measurements of turbulent fluxes exist for model comparison or validation. Early studies observed that surface snow sublimation could be significant in semiarid locations (e.g., Beaty 1975). Male and Granger (1979) showed with lysimeters and profile observations over continuous open snowfields in the Canadian prairie that turbulent fluxes often cancelled each other out because sensible heat downward to the snow drives the phase change for sublimation. Net loss of mass was smaller than 0.2 mm day^{-1} because at the Canadian prairie site, sublimation during the day was offset by condensation at night. Factors contributing to the difficulties and problems in estimating turbulent exchange from bulk transfer and flux-gradient techniques include stability and small, uncertain exchange coefficients. Typically, snow covers have low thermal conductivities and high albedos and emissivities. Because a snow surface can be very cold, especially at night, this can result in very stable conditions with a near-surface temperature inversion that will reduce turbulent mixing (Male 1980). Uncertainty in the exchange coefficients is further complicated by the in-

equality of eddy diffusivities for latent and sensible energy and momentum, and low turbulence as a result of the extreme aerodynamic smoothness of snow surfaces (Male and Granger 1979).

Currently, the most direct way to measure turbulent transfer over snow is with EC techniques (Kaimal and Finnigan 1994). Successful measurements of turbulent fluxes over snow have employed EC methodology (Harding and Pomeroy 1996; Nakai et al. 1999; Pomeroy and Essery 1999; Gryning et al. 2001; Lloyd et al. 2001; Turnipseed et al. 2002, 2003; Pomeroy et al. 2003; Molotch et al. 2007). The use of EC in more rugged terrain and within forested sites is becoming more common (e.g., Arck and Scherer 2002; Helgeson and Pomeroy 2005; Luanianen et al. 2005; Molotch et al. 2007; Pomeroy et al. 2003), however, a high degree of uncertainty is associated with these measurements. EC data over vegetated surfaces typically show energy balance (EB) closure discrepancies of 10%–30% (Harding and Pomeroy 1996; Twine et al. 2000; Wilson et al. 2002). Energy balance closure discrepancies are often attributed to differences in the measurement scale of the energy balance components, the systematic bias in instrumentation, and the loss of low- or high-frequency contributions (Wilson et al. 2002). Because melt can be used as a measure of the integrated snow cover energy balance, there are advantages to applying EC over snow. Over snow, even in complex, heterogeneously vegetated terrain or during difficult and storm conditions, the snow cover EB has been used effectively to simulate the snow cover mass balance with closure typically better than 5% (Marks et al. 2002; Garen and Marks 2005).

The objective of the research presented in this paper is to use EC measurements of heat and water fluxes from the snow cover below a forest canopy to evaluate fluxes simulated by a widely applied and validated energy balance snowmelt model. As a method, EC has been in wide use since the early 1990s, but only in the past decade or so have EC systems been applied over natural field sites with complex canopy and terrain structure. In the past few years, the reliability of EC systems has improved, and the size and power requirements have been reduced to the point that they can be operated unattended at a remote site without line power. Although the EB of the snow cover is not easily observed, the mass balance (depth, density, and melt rates) is; if the internal energetics are accounted for, they can be effectively modeled. If the EC-measured fluxes can be used to evaluate the accuracy of the simulated turbulent fluxes, then this will provide a mechanism for improving the modeling approach.

2. Site description

The research described in this paper was undertaken as part of the NASA Cold Land Processes Experiment (CLPX; Elder et al. 2009; Cline et al. 2009), conducted in Colorado during the winters of 2002 and 2003. Data used in this study were obtained during the 2003 snow season at the Local Scale Observation Site (LSOS; Hardy et al. 2008), which was within the U.S. Forest Service's (USFS) Fraser Experimental Forest (39.9°N, 105.9°W, 2780 m MSL). As part of the CLPX, snow depth and density were sampled at multiple locations within the LSOS site during two intensive observational periods (IOPs), IOP3 (late February 2003) and IOP4 (late March 2003). A micrometeorological station measuring incident solar and thermal radiation (Kip & Zonen CM-3 and CG-4), air temperature and humidity (Vaisala HMP-45), wind speed and direction (MetOne 034-B windset), soil temperature (Campbell CS-107), and snow depth (Judd Scientific sonic depth sensor) was located on the southern edge of the LSOS site within what Hardy et al. (2004) describe as a flat, uniform lodgepole pine canopy (regularly spaced tree plantation, with an average tree height of 12.6 m). The micrometeorological system was operated at the site from mid-February to the end of June 2003. An EC system was also located in this uniform pine stand and operated during the same period. The EC system included a sonic anemometer (Campbell CSAT-3) to measure the three-dimensional (3D) wind vector (u , v , w) and air temperature and an open-path infrared gas analyzer (IRGA) (Licor LI-7500) to measure water vapor at 10 Hz. The EC system also included standard slow-response instrumentation to measure 30-min average air temperature, humidity (Vaisala HMP-45), net radiation (Rebs Q7.1-L), soil temperature (Campbell CS-107), and soil heat flux (Huskeflux HFT-3.1). Unfortunately, the net radiometer was buried in a snow event in mid-March, so no reliable net radiation data were collected. Precipitation was measured at a USFS site, located approximately 100 m to the west in a small opening in a much denser forest.

3. Methods

a. SNOBAL: A two-layer energy balance snow model

In a seasonal snow cover, snow is thermodynamically unstable, undergoing continuous metamorphism until it melts (Colbeck 1982; Marks and Dozier 1992; Marks et al. 1999; Marks and Winstral 2001). These metamorphic changes and final melting are driven by temperature and vapor density gradients within the snow cover,

which are caused by heat exchange at the snow surface and at the snow–soil interface (Colbeck et al. 1979; Male and Granger 1981; Marks et al. 1999, 2002; Marks and Winstral 2001; Pomeroy et al. 1998). In general, the energy balance of a snow cover is expressed as

$$\Delta Q = R_n + H + L_v E + G + M,$$

where ΔQ is change in snow cover energy, and R_n , H , $L_v E$, G and M are net radiative, sensible, latent, conductive, and advective energy fluxes (all terms are in W m^{-2}), respectively; L_v is the latent heat of vaporization or sublimation (J kg^{-1}) and E is the mass flux by sublimation from or condensation to the snow surface ($\text{kg m}^{-2} \text{s}^{-1}$). In this context, advected energy M is heat lost or gained when mass (precipitation) of a specified temperature is added to the snow cover. In thermal equilibrium, $\Delta Q = 0.0$; a negative energy balance will cool the snow cover, increasing its cold content,¹ while a positive energy balance will warm the snow cover. The snow cover cannot be warmer than the melting temperature T_{melt} (0.0°C , or 273.16 K) and melt cannot occur until the snow, or a layer within the snow cover, has reached this temperature. Once the snow is isothermal at 0.0°C , positive values of ΔQ must result in melt.

In the next section, the model SNOBAL is briefly described. The model simulates each component of the snow cover energy balance including the turbulent fluxes (H , $L_v E$), accumulating mass and either developing a snow cover or calculating melt and surface water input or discharge from the base of the snow cover. Snow cover mass and thermal conditions are adjusted for the next time step. The model simulates the energy state and cold content of the snow cover, which is represented as a two-layer system. It predicts heat transfer, melt and liquid water drainage from each snow cover layer, discharge from the base of the snow cover, and adjusts the snow cover mass, thickness, thermal properties, and measurement heights at each time step. The modeling approach is an adaptation of that described by Marks and Dozier (1992). Although the modeling approach provides a more detailed snow cover energy and mass balance representation than that used by Wigmosta et al. (1994), Tarboton et al. (1995), or Tarboton and Luce (1996), it is a simplified form of complex multilayer models presented by Anderson (1976), Morris (1982, 1986), Flerchinger and Saxton (1989), and Jordan (1991).

The two-layer representation allows the model to ef-

¹ Cold content is the amount of energy required to bring the snow cover to T_{melt} , the melting temperature of ice (0.0°C or 273.16 K).

TABLE 1. State variables predicted by and forcing variables required by SNOBAL.

State variables	Forcing variables
Snow depth (m)	Net solar radiation (W m^{-2})
Snow density (kg m^{-3})	Incoming thermal radiation (W m^{-2})
Snow surface layer temperature ($^{\circ}\text{C}$)	Air temperature ($^{\circ}\text{C}$)
Average snow cover temperature ($^{\circ}\text{C}$)	Vapor pressure (Pa)
Average liquid water content (%)	Wind speed (m s^{-1})
	Soil temperature ($^{\circ}\text{C}$)
	Precipitation [mass (mm), temperature ($^{\circ}\text{C}$), and density (kg m^{-3})]

fectively account for the dynamic nature of snow cover energy and mass flux. It has been used to evaluate snow cover thermodynamics during rain-on-snow events (Van Heeswijk et al. 1996; Marks et al. 1998, 2001a) to assess the effects of vegetation cover on snow cover energetics (Link and Marks 1999a; Marks and Winstral 2001; Marks et al. 2001b) and to evaluate measurement techniques (Johnson and Marks 2004). By avoiding the prohibitive computational demands of more detailed multilayer representations, the spatial version of the model (ISNOBAL) has been successfully applied to predict snow cover accumulation, depletion, and stream discharge over mountain watersheds ranging in size from a few hectares to several km^2 (Link and Marks 1999b; Marks et al. 1999, 2001a,b, 2002) to regions and river basins of several thousand km^2 (Marks et al. 1999; Garen and Marks 2005) and to develop an approach for integrating snow redistribution by wind into the model (Winstral and Marks 2002; Marks et al. 2002).

Once the initial snow cover and measurement height parameters are set, the thermal, mass, and wetness conditions of the snow cover are calculated. The thickness of the lower snow layer is set as the difference between the total snow cover thickness and the thickness of the surface snow layer. The surface layer is typically modeled as a 25-cm thick layer that represents the active layer that interacts with radiation and the atmosphere. The specific mass² of each layer and the whole snow cover is calculated from layer thickness and average snow cover density. The temperature (K) of the snow surface layer and the average temperature for the entire snow cover are either set from the initial snow conditions or calculated at the end of the last model time step. The cold contents are calculated from the specific heat of ice, layer temperature, and specific mass. The specific heat of ice for each layer is calculated as a function of layer temperature. Table 1 presents the

state and forcing variables required by the model. State variables are input as initial conditions and then updated by the model during the run. Forcing variables are used by the model to predict the state variables and are input at each time step. A complete description of the model, its input requirements, and output parameters can be found in Marks et al. (1998).

b. Energy and mass balance calculations

Marks and Dozier (1992) and Marks et al. (1992) present a detailed description of equations used in SNOBAL to simulate energy and mass transfer over a snow surface, and the model structure and numerical approach are described in detail for both the point and spatial versions of the model by Marks et al. (1998, 1999, 2001a,b, 2002). Presented below is a description of the equations solved to compute sensible and latent heat fluxes, H and $L_v E$, to facilitate the comparison between EC-measured and SNOBAL-simulated turbulent transfers.

In the model, turbulent transfer terms H and $L_v E$ (W m^{-2}) are calculated using a method adapted from Brutsaert (1982) by Marks and Dozier (1992), as a system of nonlinear equations that simultaneously solve for the Obukhov stability length L (m), the friction velocity u^* (m s^{-1}), the sensible heat flux H (W m^{-2}) and the mass flux by sublimation from or condensation to the snow surface E ($\text{kg m}^{-2} \text{ s}^{-1}$):

$$L = \frac{u^{*3} \rho}{kg \left(\frac{H}{T_a C_p} + 0.61E \right)},$$

$$u^* = \frac{uk}{\ln \left(\frac{z_u - d_0}{z_0} \right) - \psi_{sm} \left(\frac{z_u}{L} \right)},$$

$$H = \frac{(T_a - T_{s,0}) a_H k u^* \rho C_p}{\ln \left(\frac{z_u - d_0}{z_0} \right) - \psi_{sh} \left(\frac{z_T}{L} \right)},$$

$$E = \frac{(q - q_{s,0}) a_E k u^* \rho}{\ln \left(\frac{z_u - d_0}{z_0} \right) - \psi_{sv} \left(\frac{z_q}{L} \right)}.$$

² Specific mass is defined as mass per unit area, or in this case mass per meter squared. For snow, it can be converted directly into a depth of water (mm) because 1 kg m^{-2} of water = 1 mm m^{-2} of water.

Here, ρ is the density of the air, k is the von Kármán constant (~ 0.40), g is the acceleration of gravity (9.81 m s^{-2}), C_p is the specific heat of dry air at constant pressure ($1005 \text{ J kg}^{-1} \text{ K}^{-1}$), E is the mass flux by sublimation from or condensation to the snow surface ($\text{kg m}^{-2} \text{ s}^{-1}$), u is the wind speed (m s^{-1}), d_0 is the zero-plane displacement height (m, $\sim (2/3) 7.35 z_0$), a_H and a_E are the ratio of eddy diffusivity for heat and water vapor to eddy viscosity, respectively. Brutsaert (1982) suggests, in the absence of other information, $a_H = a_E = 1.0$. Here, ψ_{sm} , ψ_{sh} , and ψ_{sv} are stability functions for mass, heat, and water vapor, respectively [positive when stable, negative when unstable, and 0 for neutral stability; see Brutsaert (1982) and Marks and Dozier (1992) for a detailed description of the stability functions].

The measurement heights for temperature z_T , humidity z_q , and wind z_u (m) are set as initial conditions and then updated by the model as the depth of the snow cover changes; the roughness length z_0 (m) is set as a constant at the beginning of the run but can be updated as conditions require. Air temperature T_a (K), wind speed u (m s^{-1}), and vapor pressure e_a (Pa) are model inputs, and specific humidity q (g kg^{-1}) is calculated from e_a and site air pressure. Snow surface layer temperature $T_{s,0}$ (K) is adjusted by the model at the end of each time step; snow surface specific humidity is calculated as a function of site air pressure and the saturation vapor pressure at $T_{s,0}$. The latent heat flux $L_v E$ (W m^{-2}) is $L_v \times E$, where L_v is the latent heat of vaporization or sublimation (J kg^{-1}), which varies with temperature and the state of the water (liquid or solid) from $2.501 \times 10^6 \text{ J kg}^{-1}$ for liquid water at 0°C (vaporization), or $2.834\text{--}2.839 \times 10^6 \text{ J kg}^{-1}$ for ice between 0° and -30°C (sublimation; see Byers 1974, p. 452, appendix C). A detailed description of how the model differentiates between vaporization and sublimation when liquid water is present in the snow can be found in Marks et al. (1999).

c. Eddy covariance data collection and processing

The instrumentation used at LSOS consisted of fast response (10 Hz) and mean value (30-min average) sensors. Fast response sensors included an IRGA, a 3D sonic anemometer, and a datalogger capable of collecting fast response data. Mean value sensors included a net radiometer, soil heat flux plates, soil temperature sensors, and air temperature and humidity sensors. Fast response EC data were used to calculate sensible H and latent $L_v E$ heat fluxes. At a typical EC site, mean value data on net radiation and soil heat flux are combined with the fast response data to calculate the site EB to evaluate closure. However, over snow, changes in mass

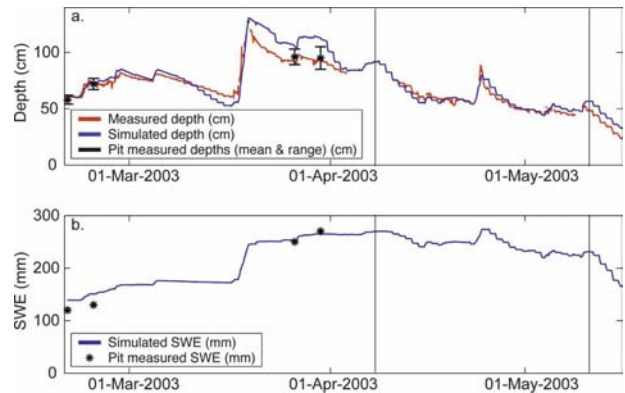


FIG. 1. Measured and simulated snow depth and SWE. Three analysis periods are indicated by vertical lines (19 Feb–7 Mar, 8 Mar–10 May, and 11–15 May). (a) Continuously measured and simulated snow depth, with manual measurements shown for the beginning and end of CLPX IOP3 and IOP4. (b) Simulated SWE with pit-measured values shown for the beginning and end of CLPX IOP3 and IOP4.

and heat storage within the snow cover were not directly measured, but modeled, for this analysis.

Data collected during the 2003 snow season were not continuous. Because of the limited data storage capacity on the EC system, there were four distinct periods of EC data collection: 16–24 February, 12–24 March, 3–17 April, and 26–29 May. These were grouped into three periods of analysis based on concurrent EC and micrometeorological data availability, snow cover, and weather conditions: the early period was cold and dry, although the snow cover doubled in mass; the mid period was a transition from cold to warmer conditions; and the late period represented active melting conditions during the final ablation of the snow cover (Fig. 1).

The eddy covariance method uses measurements of vertical fluxes in the surface boundary layer. Eddy covariance measures the flux directly by sensing the properties of eddies as they pass through a measurement level on a nearly instantaneous basis. All entities exhibit short-term fluctuations in their long-term mean values. These characteristics result in turbulence, which cause eddies to move continually around, carrying with them their properties that were derived elsewhere. The value of a given entity (s) can, therefore, be characterized by the long-term mean value $\bar{s}$ plus the instantaneous fluctuation in the value s' (Oke 1978). Therefore, $s = \bar{s} + s'$. The mean value of the entity is a time-averaged property, whereas s' is the instantaneous deviation from that mean.

An eddy is characterized by the properties contained by and transported with the eddy. These are the density ρ , the vertical velocity w , and the volumetric content of the entity s . Each of these can be broken into a long-term

TABLE 2. Modeled snow cover conditions for each of the three analysis periods. Values are for noon on the start or end day. Temperature and cold content are for the entire snow cover. Values for the start of the early period represent the initial conditions used for the model simulation.

Period	Snow depth (cm)		Snow mass (kg m ⁻²)		Temperature (°C)		Cold content (MJ m ⁻²)	
	Start	End	Start	End	Start	End	Start	End
Early (19 Feb–7 Apr)	62	91	139	269	−5.8	−4.0	−1.7	−2.2
Mid (8 Apr–10 May)	92	57	270	231	−2.8	−1.6	−1.6	−0.78
Late (11 May–15 May)	55	35	231	172	0.0	0.0	0.0	0.0

mean value and an instantaneous fluctuation. The entity can then be written as $S = (\overline{p + p'})(\overline{w + w'})(\overline{s + s'})$. The full expansion and subsequent reduction of this equation results in $S = \overline{p}(\overline{w's'})$. The overbar of $w's'$ denotes the time average of the instantaneous covariance of w and s .

Raw 10-Hz data were collected and postprocessed. For this analysis, postprocessing included statistical analysis for quality control, determination of the appropriate averaging period (Vickers and Mahrt 2003), correction for sonic temperature (Schotanus et al. 1983) and density effects (Webb et al. 1980), and tilt correction (Mahrt et al. 2000). Time series data were first processed using Quality Control (QC) Software, version 3.0 (Vickers and Mahrt 1997). The software package was designed to statistically test time series for data spikes and identify values outside of absolute limits and to assess skewness, kurtosis, discontinuities, and absolute variance. Questionable data were identified by the software and removed from the data stream. The possibility of additional sensible heat flux as a result of sensor heating (Burba et al. 2006; Grelle and Burba 2007) was also considered. However, the original equations for this correction were derived with no sensor tilt, and the author warns against the use of the equations for setups with tilted sensors. For this reason, the sensor heating corrections introduced by Burba et al. (2006) were abandoned from the data processing. Multiresolution decomposition (Howell and Mahrt 1997) was used to analyze the cospectra. A cospectral gap is seen between 8 and 10 min, which supports the use of a 10-min averaging period. The use of this averaging period should reduce the influence of mesoscale fluxes (Vickers and Mahrt 2003). The high-frequency flux loss is negligible in the latent and sensible heat fluxes because the high-frequency cospectra are near zero and substantially smaller than the peak of the cospectra. Therefore, no spectral corrections were applied to the data.

A 10-min averaging period was used to generate the covariances using second generation software (available online at http://blg.oce.orst.edu/Software/2nd_and_3rdgen/). The tilt corrected mean heat flux for each averaging period, $\overline{w'T'}$, was computed by

$$\overline{w'T'} = \overline{w'T'_v} - 0.61\theta\overline{w'q'_a},$$

where T_v is the sonic temperature (K), θ is the potential temperature (K), and q_a is the specific humidity (g g⁻¹). The sensible heat flux H is

$$H = -\rho C_p \overline{w'T'},$$

where the air density (kg m⁻³) is

$$\rho = \frac{Pm}{RT_v},$$

where C_p the specific heat of dry air (1005 J kg⁻¹ K⁻¹), P is the air pressure (Pa), m is the molecular weight of dry air (0.028 964 kg mol⁻¹), and R is the gas constant (8.314 32 J mol⁻¹ K⁻¹).

The latent heat flux $L_v E$ is

$$L_v E = -\rho L_v \overline{w'q'_a}$$

where L_v is the latent heat of vaporization and is calculated as a function of temperature. Ten-minute fluxes were then averaged over a longer time scale of one hour to reduce flux sampling errors (Vickers and Mahrt 2003) and to be compatible with the SNOBAL simulation results.

4. Results

a. Modeled versus measured snow cover energy and mass flux

The snowmelt model was initialized by measurement heights and snow cover state variables (snow depth, density, temperature, and liquid water content) based on snow cover conditions that existed at the LSOS site on 18 February 2003 (see Table 2). Although the surface layer thickness can be set, the default value of 25 cm was used for the initial simulation. The model was then driven by meteorological inputs as shown in Table 1. Net solar radiation was derived from modeled albedo (Marks et al. 1998). Simulated albedo was based on snow surface features, sun angle, and dewpoint temperature. Effective surface grain size and contamination content were modeled as a function of time and climate conditions since the last snow event (Wiscombe

and Warren 1980; Warren and Wiscombe 1980), using the broadband spectral adjustments suggested by Marshall and Warren (1987) and corrected for solar incidence angle using methods described by Dozier and Warren (1982). Spectral albedo was further adjusted for below-canopy beam and diffuse shading, using methods presented by Link and Marks (1999a,b) and for the accumulating effect of organic debris beneath the forest canopy during meltout. Forcing data were used to calculate the snow cover energy and mass balance and discharge from the base of the snow cover. The simulation was continuous over the period from 18 February to meltout on 18 May. During this period, snow depth was monitored continuously at the meteorological station, and a series of six snow pits provided periodic data on the range of depth, density, and snow water equivalent (SWE) in the surrounding area.

Fig. 1 presents a comparison of simulated and measured snow depth and SWE at the site during the simulation period. The three analysis periods listed in Table 2 are also indicated. Fig. 1a shows continuously simulated and modeled snow depth and manually sampled snow depths that were collected as part of the CLPX IOP3 and IOP4 (18–26 February 2003 and 24–29 March 2003, respectively). For the 1920 hours during which both the continuously measured and modeled snow depth were available, the RMS difference was 8.3 cm, the Nash–Sutcliffe model efficiency (Nash and Sutcliffe 1970) was 0.90, with a mean bias difference of 4.1 cm. This indicates that the simulated depths closely match measured depths but are slightly higher. A detailed description and a discussion of the appropriate use of these tests are provided by Green and Stephenson (1986), whereas the equations used and an application of the tests to SNOBAL is presented by Marks et al. (1999).

The depths from periodic snow pits and manual samples indicate both the range and the mean value for four CLPX IOP dates and show that the sonic depth measurement represents a reasonable areal average for the LSOS site. Fig. 1b compares simulated SWE (mm) to pit-measured values. Because snow pits that followed the CLPX protocols were dug only during CLPX IOP3 and IOP4, reliable measurements of SWE are limited to the accumulation and midseason phases of the LSOS snow cover. During the early and middle periods featuring dry, cold conditions, modeled snow depths and SWE matched well. Following the large storm in late March that more than doubled the amount of snow, the model tracked measured SWE fairly well but the depth was slightly overpredicted. During the initiation of warmer conditions and melt through April and into May, the model closely matched measured

depths. Although the model overpredicted the snow depth during the late March snow events, the pit data indicate that the model tracked SWE fairly well (Fig. 1b).

Table 2 presents the model results for the three analysis periods. Snow cover mass doubled during the dry, cold early period. Snow temperature and cold content both show these conditions, although density increased from 225 to 295 kg m⁻³. Little melt and no discharge were generated during the early period. During the mid period transition from cold to warmer conditions, the snow temperature increased and cold content decreased, with densities rising from 297 kg m⁻³ to 412 kg m⁻³. Some melt occurred, with the loss of about half the depth but only 15% of the mass. A mass of 40 mm was lost to discharge and sublimation, whereas much of the melt was retained and refrozen, resulting in higher densities and shallower depths. During the late period's³ four days of active melting, the snow was at or near the melting temperature, and the cold content was negligible. A steady decrease in depth is shown in both the simulated and measured values, with a 20-cm loss of depth, a 58-mm loss of mass, and a change in density from 416 to 498 kg m⁻³.

Fig. 2 shows the SNOBAL-simulated energy and mass fluxes for the 86 days (2064 h) of analysis during the 2003 snow season. During the early period, daily values of ΔQ were near zero, with positive daytime values balanced by negative values at night and with the trend moving to greater daytime positive values during the mid period and then to positive values during both day and night during late period meltout conditions (Fig. 2a). Net radiation R_n generally mirrored ΔQ , beginning in balance between negative and positive values, moving toward increasing daytime positive values, and finally to positive values during both day and night during meltout (Fig. 2b). Sensible H and latent heat $L_v E$ show the typical oversnow balance reported by Marks et al. (1998, 1999, 2001a) between positive H and negative $L_v E$, as the snow cover was warmed by sensible heat and cooled by sublimation. During the early and mid periods, this balance was biased toward sublimation; however, during meltout, sensible heat to the snow cover began to dominate over latent heat losses (Fig. 2c). Sublimation was continuous during the snow season, with peaks between storm periods and lows during precipitation events. In general, sublimation was greatest during cold conditions but diminished as con-

³ Precipitation data were obtained from a nearby USFS station, which ceased operations on 5 May. Analysis was terminated after 15 May because a rain-on-snow event occurred for which no precipitation data were available.

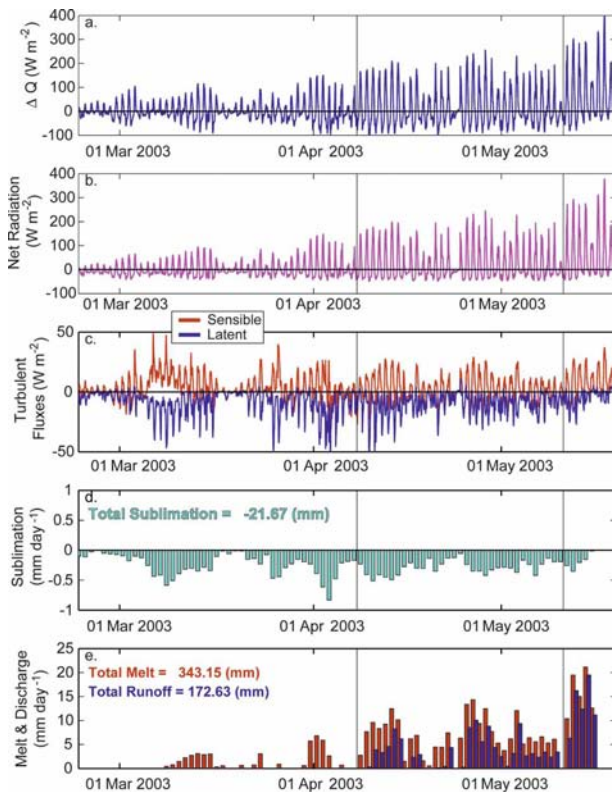


FIG. 2. Simulated mass and energy balance components. The three analysis periods indicated by vertical lines.

ditions warmed and humidity increased (Fig. 2d). Melt began in March, but no discharge was generated until mid-April. As the snow cover warmed and conditions favored snow cover “ripening” (snow temperature = T_{melt} , with liquid water saturation), greater portions of melt percolated through and were lost from the snow cover as discharge. By the late period, with conditions ideal for rapid meltout, most melt translated directly into discharge.

Note that, as shown in Table 3, cumulative melt (343) minus discharge (173) equals 171 mm, which exceeds

the ending model-specified water holding capacity of the snow cover (5 mm) and the theoretical wet snow cover limit (22 mm) by 151–166 mm. This “excess” meltwater is refrozen and remelted many times during the diurnal cycling of melting and freezing conditions that are typical of a seasonal snow cover (Fig. 2e). Only liquid water that exceeds the specified holding capacity of the snow cover is translated by the model into discharge. Although there is some uncertainty about this value (Davis et al. 1985; Denoth et al. 1984), most evidence supports the conclusion that in the absence of ice layering, it seldom exceeds 1% of the snow cover void space; although in the presence of ice layering in a wet, melting snow cover it can be as much as 5%. As conditions cool during the night, negative ΔQ is used to refreeze meltwater retained by the snow, increasing the density and reducing the cooling effect on the snow cover. During the early period, no discharge was generated and nearly all of the melt was refrozen. Though some discharge was generated during the mid period, much of the melt was also refrozen. By the late period nighttime refreezing was no longer occurring, and melt was being converted directly into discharge.

Table 3 presents a summary of simulated snow cover energy and mass fluxes during the three analysis periods, as shown in Fig. 2. Although the early period was nearly in balance with ΔQ near zero, the EB was dominated by soil heat and turbulent fluxes. Although the soil heat flux G was not large, it was important because the soil was significantly warmer than the snow. Here, R_n was near zero, with thermal and solar radiation canceling each other. During the mid period, as solar radiation increased with higher sun angles, net radiation R_n replaced soil heat and turbulent fluxes G and $H + L_v E$, respectively, as the dominant energy balance processes. As the snow cover warms, the temperature gradient between the soil and snow is reduced, as are the temperature and moisture gradients between the atmosphere and the snow. During the late period, the snow cover EB became a radiation-dominated system. Tur-

TABLE 3. Snow cover energy and mass flux summary. Average simulated energy and total mass flux for the three analysis periods and for the whole simulation period. Evaporation is shown as a negative value because the flux is to the atmosphere; condensation is considered a positive flux. Note that 1) because energy fluxes are averaged over the analysis periods, they generally represent a wide range of values; 2) total evaporation includes condensation; 3) total melt is not adjusted for refreezing; and 4) total discharge includes both meltwater and rain.

Period	Average energy flux (W m^{-2})							Total mass flux (mm)		
	ΔQ	R_n	H	$L_v E$	$H + L_v E$	G	M	Evap	Melt	Discharge
Early	0	−1	4	−8	−4	5	0	−11	45	0
Mid	14	20	2	−10	−8	2	0	−10	219	107
Late	54	53	7	−7	0	1	0	−1	79	66
All	9	10	3	−9	−5	3	0	−22	343	173

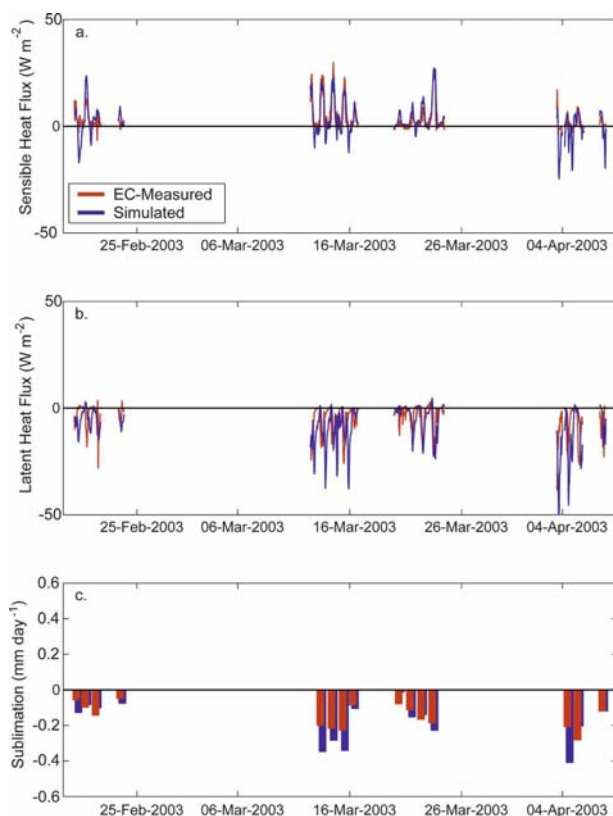


FIG. 3. Early period EC-measured (red) and SNOBAL-simulated (blue) (a) sensible heat flux, (b) latent heat flux, and (c) sublimation.

bulent fluxes balanced each other, and the soil and advective fluxes were small, so that R_n was converted almost directly to energy for melt.

Soil heat flux was consistently small during the snow season, with a decrease in G to about 1.5 W m^{-2} after the cold early period. Advected heat flux M from precipitation was insignificant throughout the period of analysis. Latent heat flux $L_v E$ was greater in magnitude than sensible H for most of the snow season but went into balance with H during the final meltout. During the early period, only a small amount of melt and no discharge were generated; however, during the mid period a significant amount of melt was generated, with roughly half of that converted to discharge. Significant melt was also generated during the late period, with nearly 85% converted to discharge. Sublimation was consistent throughout the snow season at about $-0.25 \text{ mm day}^{-1}$, which represents around 22 mm of SWE, or a loss of 6.5% of the snow mass during the 86 days (2064 h) from 19 February to 15 May 2003. This value is higher than those reported by Male and Granger (1979) and Pomeroy et al. (1998) for the Canadian prairie and much higher than reported by Harding and

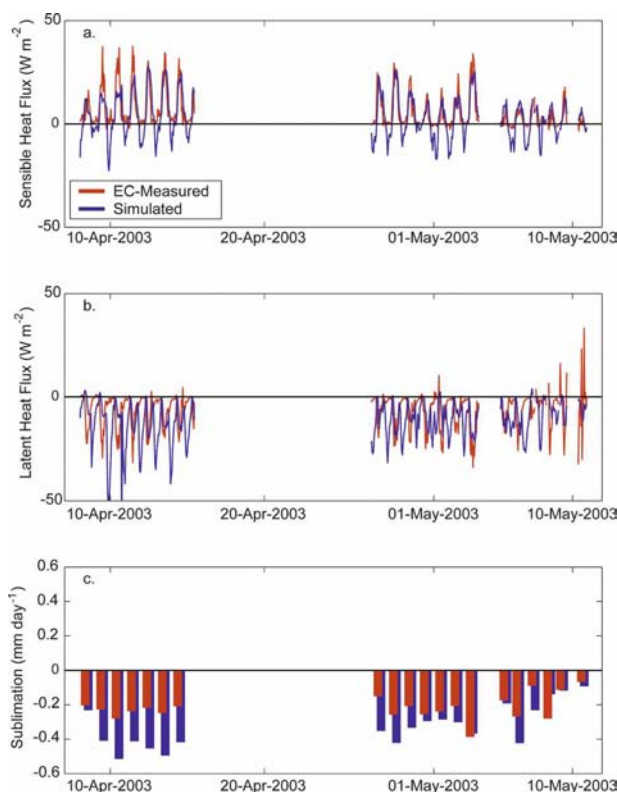


FIG. 4. Same as Fig. 3 but for mid period.

Pomeroy (1996) for snow under boreal forest canopies. However, at these more northerly and generally stable sites, we would expect much less radiative energy for sublimation and turbulent transport than at midcontinent subalpine site such as the Fraser Experimental Forest.

b. Modeled and EC-measured turbulent energy and mass components

During the 86 days of the EC and model analysis, only 39 days (927 h) of data were viable for comparison. Although the early period spanned 48 days because of precipitation events, only 15 days (355 h) of data were viable for comparison (Fig. 3). During this period, EC and modeled sensible heat flux H compared well (Fig. 3a), but modeled latent heat flux $L_v E$ was greater in magnitude than EC $L_v E$, particularly at night (Fig. 3b). When latent heat was converted to daily mass loss, EC sublimation was also less than modeled evaporation (Fig. 3c). During the 33 days (792 h) of the mid period, there were fewer precipitation events, so 19 days (462 h) of viable data were available for comparison (Fig. 4). As for the early period, EC and modeled sensible heat flux H compare well, especially during the day (Fig. 4a). Fig. 4b shows that the 8–14 April segment was also

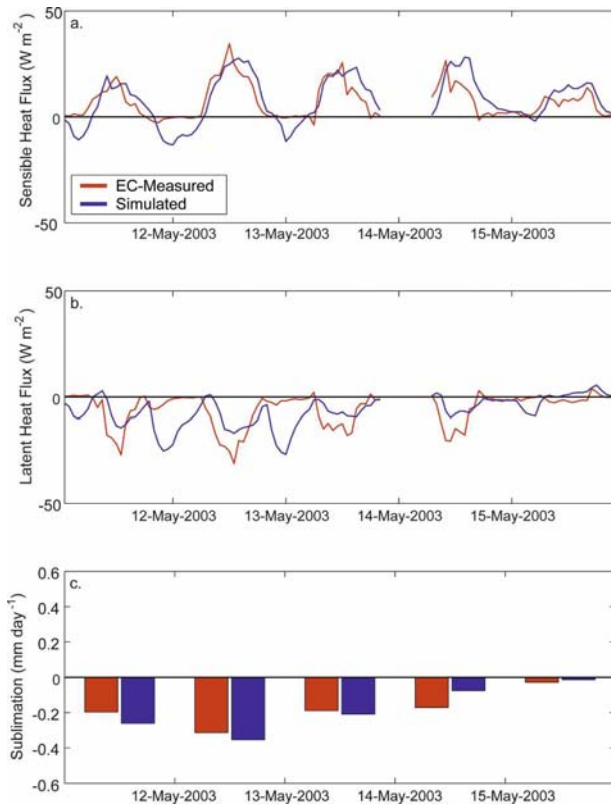


FIG. 5. Same as Fig. 3 but for late period. In (c), each pair of blue and red bars represents EC-measured and SNOBAL-simulated sublimation totals for a single day.

similar to the early period, showing a modeled latent heat flux $L_v E$ greater in magnitude—particularly at night—than EC $L_v E$. However, data from 26 April–10 May showed comparable magnitudes for both modeled and EC $L_v E$, though the nighttime discrepancy was still evident. Similarly, the 8–14 April modeled daily sublimation was greater than EC observations, but during 26 April–10 May the modeled and EC sublimation are similar (Fig. 4c). The late period of active meltout conditions was only five days, but nearly all were viable for comparison (Fig. 5). Both modeled sensible and latent heat fluxes (H , $L_v E$) compare well to EC H and $L_v E$ during this period, though there are differences during night (Figs. 5a, b). The daily modeled and EC sublimation were nearly equivalent during this period (Fig. 5c).

Table 4 is a summary of the modeled and EC observations expressed as turbulent fluxes (energy) and sublimation (mass). Fluxes are very small in all cases. Simulated sensible heat flux H generally compares well to EC-measured H . Differences were less than 1 W m^{-2} during both early and late periods (mean values 3–4 and $7\text{--}8 \text{ W m}^{-2}$, respectively), although they are slightly larger (3 W m^{-2}) during the mid period (mean values

TABLE 4. Summary of EC-measured and SNOBAL-simulated average sensible and latent heat flux (H and $L_v E$) and total evaporation (mm) for the periods when viable concurrent values were available during each of the three analysis periods (see Figs. 3–5). The number of hours of concurrent data is indicated for each period and for the entire record.

Period	Sensible heat flux $H \text{ (W m}^{-2}\text{)}$		Latent heat flux $L_v E \text{ (W m}^{-2}\text{)}$		Evaporation (mm)	
	EC	SNOBAL	EC	SNOBAL	EC	SNOBAL
Early (355 h)	4	3	−5	−8	−3	−3
Mid (462 h)	6	3	−7	−11	−4	−7
Late (110 h)	7	8	−6	−7	−1	−1
All (927 h)	5	4	−6	−10	−8	−11

$3\text{--}6 \text{ W m}^{-2}$). Overall, the difference was about 1 W m^{-2} , with the EC-measured H being slightly higher. The magnitude of simulated latent heat flux $L_v E$ was greater over all the analysis periods than the EC-measured $L_v E$, showing a difference of 4 W m^{-2} (mean values from -6 to -10 W m^{-2} overall). Differences were greatest during the mid period (4 W m^{-2}) but were small during late period melting conditions. Differences in latent heat flux estimation resulted in cumulative sublimation differences of 3 mm or 0.08 mm day^{-1} . Although this is a very small difference, it is nearly 31% of the 0.25 mm daily value simulated by the model.

5. Discussion

The EC-measured and SNOBAL-simulated turbulent fluxes (H and $L_v E$, respectively) were of similar direction and magnitude, although EC-measured fluxes almost perfectly balanced with each other, whereas the SNOBAL-simulated fluxes tend to more negative values. EC and SNOBAL sensible heat flux H tended to agree more closely than latent heat flux $L_v E$ and, hence, sublimation. For instance, SNOBAL estimates of $L_v E$ exceeded EC-measured $L_v E$ by 31% but because the fluxes were small, the cumulative differences were very small (1 W m^{-2} for H , 4 W m^{-2} for $L_v E$, and 3 mm for sublimation) over the 927 h of data used for comparison. If the EC energy flux terms are assumed correct, complete snow depletion (meltout) would have occurred about $3\frac{1}{2}$ days sooner than the model prediction. Actual meltout is difficult to determine because of large energy inputs from the mixed-rain-and-snow event during 16–18 May and because the snow cover became patchy after 18 May. The depth sensor indicated meltout occurred on 21 May, whereas the results from the EC system indicated that it occurred on 18

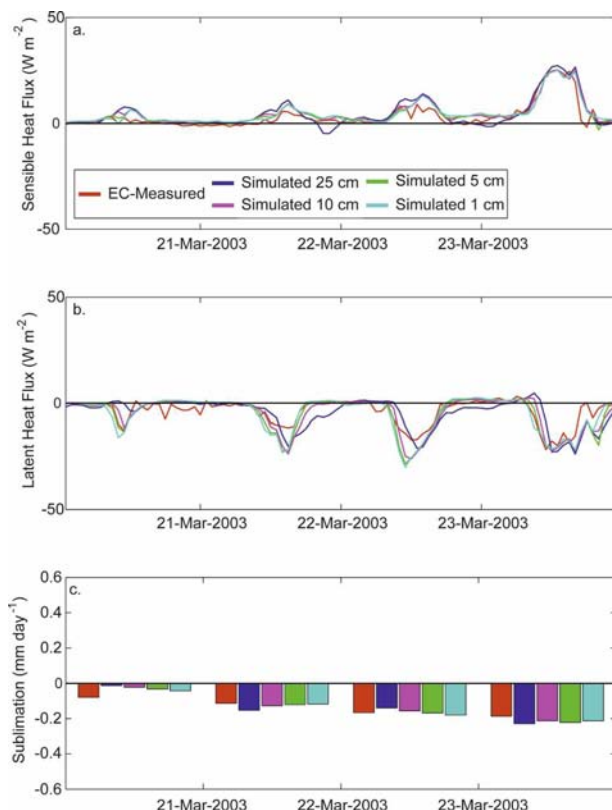


FIG. 6. Early period EC-measured and SNOBAL-simulated (a) sensible, (b) latent heat flux, and (c) sublimation with varying model surface layer thickness. In (c), each grouping of bars represents measured and modeled total sublimation for a single day.

May following the storm; and the SNOBAL-simulated meltout occurred on 23 May. The difference between EC-measured and SNOBAL-simulated sublimation is so small (0.08 mm day^{-1} over the 39 days of comparison) that we consider it to be within the noise of both systems. Although the differences between EC-measured and SNOBAL-simulated fluxes are small, two aspects of the differences are of interest and will be discussed below.

a. Model snow surface layer thickness

The first issue of interest in comparison between EC-measured and SNOBAL-simulated fluxes is the substantial discrepancy that occurs at night when EC-measured fluxes are close to zero but simulated H shifts to negative and $L_v E$ becomes more negative. This occurs in the model because the snow surface layer is warm and has thermal inertia that must be overcome to cool the entire layer before the temperature and moisture gradients are eliminated. The EC-measured fluxes appear to respond to a much thinner surface layer that rapidly cools as the sun goes down (Figs. 3a,b and 4a,b).

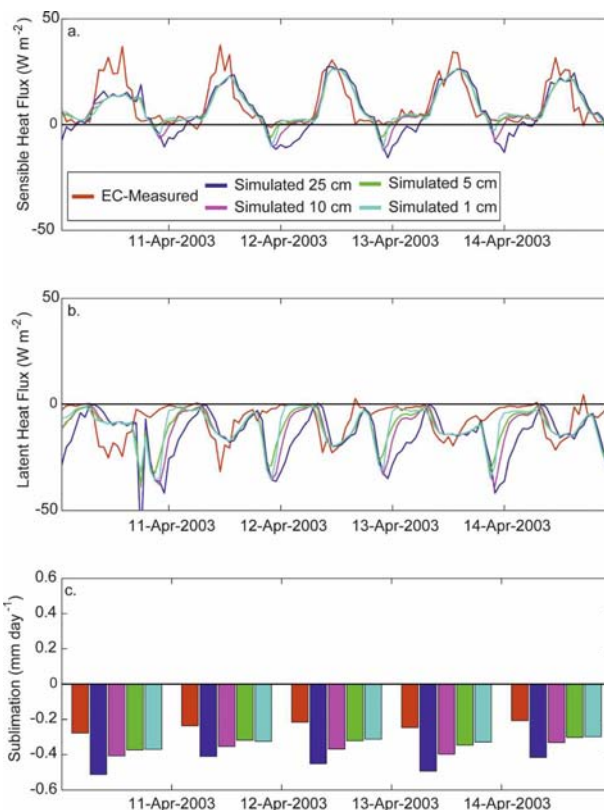


FIG. 7. Same as Fig. 6 but for mid period.

Figures 5a and 5b show that this effect is reduced but still evident as the snow cover warms. The snow is a porous medium, and turbulent fluxes are normally considered to be a result of temperature conditions some distance into the snow cover (Andreas 1987; Jordan et al. 1999), although there is uncertainty about the effective thickness of the exchange layer within the snow cover. It can be assumed that the EC-measured fluxes were responding to the true exchange layer, whereas the model exchange layer thickness is arbitrary. If differences between simulated and measured fluxes are sensitive to changes in the model-specified surface exchange layer, then as the model surface layer is reduced, we would expect the simulated fluxes to more closely match the EC-measured fluxes. To test the sensitivity of the simulation to the specified surface layer thicknesses, the model was rerun using thicknesses of 10, 5, and 1 cm.

Figures 6, 7, and 8 present a comparison of EC-measured and SNOBAL-simulated turbulent fluxes from these simulations for selected dates during each of the three analysis periods: Fig. 6 for $4\frac{1}{2}$ days, 20–25 March, during the early period; Fig. 7 for 5 days, 10–14 April, during the mid period; and Fig. 8 for 5 days,

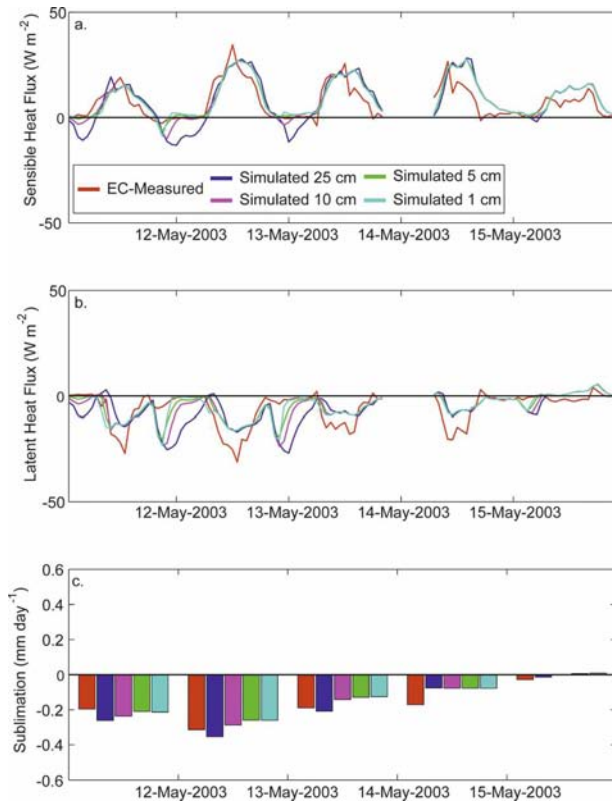


FIG. 8. Same as Fig. 6 but for late period.

11–15 May, during the late period. During the 39 days of comparison, the effect on sensible heat flux H was small, but as the model surface layer thickness was made thinner, differences between EC and modeled latent heat flux $L_v E$ and, therefore on sublimation flux

differences, were reduced. Table 5 presents a summary of EC-measured turbulent fluxes compared to SNOBAL-simulated fluxes for model surface layer thicknesses of 25, 10, 5, and 1 cm for each of the three analysis periods and for the entire EC–SNOBAL model comparison period. Figures 6, 7, and 8 and Table 5 show that specification of a thinner surface layer in SNOBAL substantially reduces or eliminates differences between EC-measured and SNOBAL-simulated turbulent fluxes. By reducing the model surface layer thickness, the differences between EC and modeled latent heat flux $L_v E$ and, therefore on sublimation flux differences, were reduced from 4 to 2 W m^{-2} for $L_v E$ and from 3 to 1 mm (0.02 mm day^{-1}) for evaporation over the 39 days of comparison. They also show that most or all of the improvement occurs with an active layer specification of 10 cm and that a specification of 5 or 1 cm thickness for the active layer has only a small impact on the result.

b. Daytime-only fluxes

The second issue of interest in comparison between EC-measured and SNOBAL-simulated turbulent fluxes is the concern that EC measurements and simulations are more difficult during the night as a result of very stable air temperature stratification, low heat fluxes, and low wind velocities (Foken and Wichura 1996). Because of variable snow depths, the EC system was located on a tower approximately 3.5 m above ground level. This was a compromise between locating it in the below-canopy region (<2.5 m above the ground) and keeping it 2 m above the snow. There is a potential for the EC system to fail to measure the effect

TABLE 5. Summary of EC-measured and SNOBAL-simulated average sensible and latent heat flux (H and $L_v E$) and total evaporation (mm) for surface layer thickness of 25, 10, 5, and 1 cm. (see Figs. 6–8).

Period	Sensible heat flux H (W m^{-2})				
	EC	SNOBAL–25 cm	SNOBAL–10 cm	SNOBAL–5 cm	SNOBAL–1 cm
Early (355 h)	4	3	4	4	4
Mid (462 h)	6	3	5	5	5
Late (110 h)	7	8	9	9	9
All (927 h)	5	4	5	5	5
Period	Latent heat flux $L_v E$ (W m^{-2})				
	EC	SNOBAL–25 cm	SNOBAL–10 cm	SNOBAL–5 cm	SNOBAL–1 cm
Early (355 h)	–5	–8	–8	–8	–8
Mid (462 h)	–7	–11	–9	–8	–8
Late (110 h)	–6	–7	–5	–5	–5
All (927 h)	–6	–10	–8	–8	–8
Period	Evaporation (mm)				
	EC	SNOBAL–25 cm	SNOBAL–10 cm	SNOBAL–5 cm	SNOBAL–1 cm
Early (355 h)	–3	–3	–3	–3	–3
Mid (462 h)	–5	–7	–5	–5	–5
Late (110 h)	–1	–1	–1	–1	–1
All (927 h)	–8	–11	–9	–9	–9

TABLE 6. Summary of EC-measured and SNOBAL-simulated average sensible and latent heat flux (H , $L_v E$) and total evaporation (mm) for daylight hours (in which measured incoming solar radiation was greater than zero) for hours when viable concurrent values were available during each of the three analysis periods. Modeled values are presented for both 25 and 10 cm surface layer thicknesses. The number of hours of concurrent daylight data is indicated for each period and for the entire record.

Period	Sensible heat flux H (W m^{-2})			Latent heat flux $L_v E$ (W m^{-2})			Evaporation (mm)		
	EC	SNOBAL		EC	SNOBAL		EC	SNOBAL	
		25 cm	10 cm		25 cm	10 cm		25 cm	10 cm
Early (204 h)	6	6	5	-8	-8	-10	-2	-2	-2
Mid (274 h)	10	10	9	-10	-8	-8	-4	-3	-3
Late (74 h)	10	13	13	-8	-5	-5	-1	-1	-1
All (552 h)	8	9	8	-9	-8	-8	-7	-5	-6

of very large nighttime temperature and moisture gradients at the snow surface on turbulent fluxes. To test this, daytime-only fluxes were extracted from the analysis shown in Figs. 6–8 for reanalysis. Daytime was defined as solar radiation greater than zero. This reduced the number of hours of viable data for concurrent analysis from 927 to 552. Table 6 presents a summary of the daytime-only comparison between EC-measured and SNOBAL-simulated turbulent fluxes. In general, the daytime-only comparison shows an almost exact agreement between EC-measured and SNOBAL-simulated turbulent fluxes. This also indicates that specification of surface layer thickness does not substantially influence this result. During the analysis period, the 1 W m^{-2} difference between EC-measured and SNOBAL-simulated sensible heat flux H was unchanged, but the latent heat flux $L_v E$ difference was reduced from 4 to 1 W m^{-2} , and the sublimation difference was reduced from 3 to 1 mm.

6. Conclusions

At the LSOS site, EC-measured and SNOBAL-simulated turbulent fluxes agree during periods of viable concurrent data within the 86-day analysis period when appropriate snow cover exchange layers were included in the model parameter set. Although the measured and simulated fluxes show small differences during the analysis period, the largest discrepancies occurred at night. During night, conditions are stable with very cold near-surface temperatures and the dominance of longwave radiation exchange at the snow surface. Under these conditions, large temperature and moisture gradients may exist near the snow surface. Observations have shown that in low-turbulence snow environments such as the LSOS site, the presence of a shadow can reduce the radiant surface temperature by 8° – 10°C (Rowlands et al. 2002). Once the sun goes down, large snow surface temperature and moisture

gradients will occur and persist until the near-surface layer has also cooled. If we reduce the thickness of the snow surface exchange layer in the model, this equilibrium occurs more quickly, and the EC-measured fluxes more closely match the SNOBAL-simulated fluxes.

While using SNOBAL to test a range of surface layer thicknesses at the LSOS site, we find that there is considerable decrease in difference between EC-measured and SNOBAL-simulated fluxes when surface layer thickness is reduced from 25 to 10 cm but that further reduction to 5 or 1 cm does not improve the results. In SNOBAL, the default surface layer thickness of 25 cm was established because that was the assumed depth of radiation penetration into snow (e.g., Nolin and Dozier 1993). However, solar radiation is not a factor in nighttime energy balance, and the thickness of the snow cover exchange layer is more likely a result of turbulent transfer and conduction between the atmosphere and a radiatively cooling snow surface. Both models and measurements show that turbulent fluxes occur across a snow surface layer of varying thickness, depending on snow density, wind, and atmospheric pressure fluctuations (Marsh et al. 1997; Jordan et al. 1999; Essery et al. 2003; Massman 2006; Massman and Frank 2006). Based on the analysis presented here, it would appear that the active layer thickness for turbulent transfer over the LSOS snow cover was equal to or less than 10 cm.

Nighttime stability and instrument placement resulted in some uncertainty in the EC-measured fluxes. During the 927-h comparison period when viable EC data were available, SNOBAL-simulated $L_v E$ was 4 W m^{-2} more negative and sublimation was 3 mm greater than EC-measured values. Comparing only daytime values, EC and SNOBAL fluxes (H , $L_v E$, and evaporation) were equivalent. This suggests that either large surface temperature and moisture gradients that occur over snow during very stable conditions are not well sensed by EC technology, or they are not well simulated by the homogeneous layer approach employed in

SNOBAL. It also indicates that model surface layer thickness is a parameter that must be carefully selected, depending on radiation, wind, and snow structure conditions.

Both point and spatial versions of the snow model have been successfully applied and extensively validated at a variety of sites and under a range of conditions, using the default 25-cm surface layer thickness (Garen and Marks 2005; Marks and Winstral 2001; Marks et al. 1998, 1999, 2001a,b, 2002), although most of these applications were over relatively warm snow conditions. However, in the model applications over a forested snow cover in boreal Canada by Link and Marks (1999a,b), model performance improved once the specified surface layer thicknesses were reduced. At the time, the adjustment was made because of very cold conditions and a thin snow cover. No winter flux data were available for the boreal site, but observed turbulent conditions beneath the boreal forest were different from those at the LSOS site. Although both the LSOS and boreal snow covers were cold and the air was dry, there was much less wind at the boreal site, so near-surface conditions were very stable. Further investigation of the appropriate exchange layer thickness for snow cover energetics is warranted. The data presented in this paper are limited to low-turbulence, below-canopy conditions. Without similar EC measurements, and meteorological and snow data at an open, wind-exposed site, we cannot determine if the same day/night and surface layer thickness concerns apply to other than below-canopy sites.

In this experiment, we compare simulated and EC-measured fluxes of heat and water from the snow surface, with the objective of providing an evaluation of methods used by the model to simulate turbulent fluxes that could lead to an improved modeling approach. This was a limited experiment; although careful processing and correction of the EC data was performed, a critical evaluation of the EC technology used or uncertainties in the EC-measured fluxes was not undertaken. Although we treat EC as a measurement technology, it more realistically provides a model-based estimate of fluxes that can be quite sensitive to a complex suite of site and conditional assumptions. How well these assumptions were satisfied at the LSOS below-canopy site was beyond the scope of this experiment.

The experiment demonstrates the use of eddy covariance methods for measuring heat and mass fluxes from snow covers at a midlatitude, subcanopy site and shows that in addition to traditional mass balance methods, EC-measured fluxes can be used to diagnose the performance of a snow cover energy balance model. This research resulted in an improved model surface layer

thickness configuration and a better understanding of the sensitivity of the snow cover surface exchange layer thickness to temperature and stability conditions. Nearly exact agreement between simulated and EC-measured fluxes occurred when the adjusted surface layer thickness was applied and when comparisons were limited to daylight hours. The application of the adjusted surface layer thickness in the model improved nighttime comparisons but did not resolve all the differences between the simulated and EC-measured fluxes.

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