Modelling snow–atmosphere interactions in cold continental environments

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Abstract Land surface process schemes are very sensitive to snow cover and snow processes. However many of these schemes are deficient in their representation of snow processes, as manifested in cold continental climates. The next generation of GCMs, meso-scale models and coupled hydrological models will require improved representations of snow surface processes. To this end, new algorithms have been developed and verified by field measurements to improve snow process representations for prairie and arctic environments. Based on the algorithm results, recommendations for advances in snow modelling are made regarding: redistribution of snow cover by blowing snow, sublimation loss during blowing snow, turbulent transfer during snowmelt and ground heat flux during percolation of snowmelt water through the snowpack and infiltration into frozen soil.

INTRODUCTION

Algorithms that describe snow processes are important components of land surface schemes for General Circulation Models (GCM) because much of the Earth's land surface is covered with snow and ice and strong snow–climate feedbacks have been identified from model output (Randall *et al.*, 1994) and measurements (Karl *et al.*, 1993). A recent review of snow accumulation and ablation processes in land surface schemes has identified several key deficiencies in present schemes, prominent amongst which are the lack of algorithms for blowing snow redistribution and sublimation, overestimation of net turbulent transfer to continuous snow covers and inadequate representation of heat flux during snowmelt infiltration to unsaturated frozen soils (Pomeroy *et al.*, 1998). The objectives of this paper are to examine the importance of these processes in defining atmospheric fluxes during snow accumulation and melt sequences in cold open environments and to recommend improvements.

EXPERIMENTAL SITES AND DATA COLLECTION

Data was collected on a level, well-exposed field (Kernen Farm) near Saskatoon in the northern prairies of western Canada. Saskatoon has a northern continental climate with daily winter temperatures ranging from just above 0°C down to approximately –40°C and relatively-low annual snowfall. Melt occurs in March or April. Meteorological data was collected from instrumentation (radiometers, hygrothermometers, ultrasonic anemometers) designed to measure components of the energy balance.

PROCESSES

Snow transport and sublimation by wind

Redistribution of snow by wind relocates snow covers and produces notable in transit sublimation of blowing snow (Dyunin, 1959; Schmidt, 1972; Pomeroy, 1989). Relocation involves a horizontal mass flux of snow proportional to the fourth power of wind speed, and sublimation a vertical flux of water vapour proportional to the fifth power of wind speed, depending on fetch. The snow accumulation flux, Q_A (kg m^{-2} s^{-1}), over some fetch distance, F (m), during blowing snow may be described as,

$$Q_A(F) = P - \frac{Q_R(F) - Q_R(0)}{F} - Q_E \tag{1}$$

where, P is snowfall (kg m^{-2} s^{-1}), Q_R is downwind blowing snow transport (kg m^{-1} s^{-1}) and Q_E is sublimation (kg m^{-2} s^{-1}). As evident from equation (1), at the large spatial scales (fetches) of GCM grid cells, sublimation becomes the most important blowing snow flux, though transport strongly affects sub-grid variability (smaller fetches) in snow accumulation. Reported annual fluxes of blowing snow sublimation range from 15% to over 40% (depending on climate, fetch and land use) of annual snowfall on the Canadian Prairies (Pomeroy & Gray, 1995), 28% of annual snowfall on tundra in the Western Canadian Arctic (Pomeroy *et al.*, 1997) and 32% of annual snowfall on the Alaska north slope (Benson, 1982).

Land surface schemes do not presently incorporate blowing snow processes, however the Prairie Blowing Snow Model (Pomeroy, 1989) has been redeveloped for compatibility with GCMs, incorporating features for independent determination of the transport threshold for drifting, upscaling blowing snow fluxes using probability theory, improved vegetation parameterizations, simplified calculations for variable fetches and landscape-based snow mass balances that include snowmelt (Pomeroy *et al.*, 1997; Pomeroy & Li, 1997) and flow over complex terrain (Essery *et al.*, 1999).

Snowmelt energetics

Land surface schemes calculate the energy available for snowmelt, Q_m, by an energy balance equation. Therefore, assuming a continuous snow cover:

$$Q_m + Q_n + Q_h + Q_e + Q_a + Q_g = \mathrm{d}U/\mathrm{d}t \tag{2}$$

The terms of equation (2) are: Q_n, net radiation, Q_h, turbulent flux of sensible heat exchanged at the surface due to a difference in temperature between the surface and overlying air, Q_e, turbulent flux of latent energy exchanged at the surface due to vapour movement as a result of a difference in vapour pressure between the surface and overlying air, Q_a, energy advected to the snowpack from rainfall or other processes, Q_g, ground heat flux due to conduction, U, internal energy, and t, time. Discussion of the Q_a term is provided by Marsh *et al.* (1999). Comments on appropriate snow albedo for net radiation calculation are provided by Pomeroy *et al.* (1998).

Turbulent fluxes Land-surface models parameterize exchanges of heat and moisture between the atmosphere and the surface using Dalton-type bulk transfer relationships. Surface humidity is generally assumed to be saturated over snow and the surface temperature is calculated from the energy balance but cannot rise above 0°C; snowmelt is diagnosed from the energy required to balance the surface energy budget subject to the surface temperature constraint. Exchange coefficients are calculated as functions of surface roughness and some index of atmospheric stability—either a Monin-Obukhov length or a Richardson number. Factors contributing to the difficulties and problems in calculating turbulent exchange in this manner include the validity of the assumption of a constant flux layer, extremely stable conditions of the atmospheric surface layer that dampen turbulent mixing, unequal eddy diffusivities for latent and sensible energy and momentum and low roughness lengths compared to other earth surfaces (Male, 1980; Male & Granger, 1979).

Ground heat and internal snowpack energy Most simulations of ground heat in land surface models are based on heat transfer by conduction using the temperature gradient approach and simulated snow/soil temperatures at three to five levels in a profile that extends to the rooting depth of the crop or below. Often, heat transfers due to phase changes from freezing and thawing are either ignored or the freezing point depression relationship—the curve describing the association between liquid water content and freezing temperature—is represented as a step function rather than as a continuous curve.

When meltwater infiltrates a cold snow cover, water may freeze as ice layers within the snowpack, resulting in warming of both the snow and underlying soil by conduction of the released latent heat of fusion (Marsh & Woo, 1984a). Once meltwater reaches the base of the snow cover it may be unable to infiltrate the frozen soil or the infiltration capacity may be very small compared to the meltwater flux. These conditions can occur when an ice layer has formed above the soil surface or for uncracked, ice-rich permafrost soils. If the soils are very cold and the soil infiltration capacity is small, the combination of a saturated layer at the base of the snowpack, along with intense temperature gradients and ground heat fluxes of up to 80 W m^{-2} (Marsh & Woo, 1987), lead to the formation of basal ice layers at the soil surface (Woo *et al.*, 1982). This refreezing results in rapid increases in soil temperature and decline in ground heat flux (Marsh & Woo, 1984b). Many of these changes are not normally accounted for by temperature gradient methods as used in snowpack/soil models which often have a limited number of layers, as is common in land surface schemes.

For most northern regions, some infiltration of meltwater into frozen soils accompanies snow ablation. Under these conditions the ground heat flux is usually small (<5 W m^{-2}) compared to the melt energy flux and the important heat and mass transfer processes affecting the flux occur in the upper 30 cm of the soil temperature profile. Field measurements (Kane & Stein, 1983) and model simulations (Zhao *et al.*, 1997) demonstrate that both the infiltration rate and the surface heat transfer rate (conduction) decrease rapidly with time following the application of meltwater to the surface (Fig. 1). Zhao *et al.* (1997) suggested that these variations may be described by two regimes, a transient regime and a quasi-steady state regime. The transient regime immediately follows the percolation of meltwater to the soil surface and persists for a few hours, during which the energy used to increase the soil temperature is largely supplied by heat conduction at the surface. The quasi-steady state regime develops when the changes in the infiltration rate and the heat transfer rate with time become relatively small. In the quasi-steady state regime, the energy used to increase soil temperature at depth is supplied by latent heat released by the refreezing of infiltrating meltwater in the soil layers above (Zhao *et al.*, 1997).

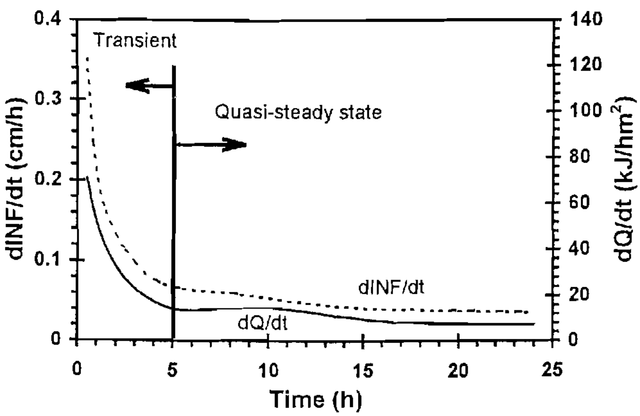

Fig. 1 Variations in infiltration rate (d*INF*/d*t*) and surface heat flux rate (d*Q*/d*t*) with time during snowmelt infiltration into a frozen silty-clay soil.

FIELD EVALUATIONS OF MODELS

Latent heat flux during blowing snow

The latent heat flux due to sublimation is of interest to both hydrologists and meteorologists. An evaluation of the Prairie Blowing Snow Model calculation of latent heat flux due to blowing snow, and that calculated using a Dalton-type bulk transfer formulation with a fixed roughness height, was conducted over an open snowfield near Saskatoon in December 1998 and January 1999 using measurements from an eddy correlation system. For roughness length modelling, 1 cm of exposed vegetation above the snowpack was presumed. The blowing snow model used measured air temperature and humidity (2 m) and wind speed (2 m); the upper boundary layer for sublimation calculations was artificially set to 1 m which roughly corresponded to the greatest height for generation of water vapour-containing eddies that could be detected by the eddy correlation system (at 2 m). The bulk transfer latent heat calculation used measured snow surface and air temperature (2 m), measured humidity at 2 m, measured wind speed and a roughness length of 1 mm. The effective roughness length for water vapour transfer was considered 0.1 of that for momentum as is standard practise in land surface schemes.

Meteorological conditions during the two blowing snow events and adjacent non-blowing periods (Day 351–353, 1998 and Day 365, 1998–2, 1999) are shown in Fig. 2(a). The first blowing snow storm involved wind speeds exceeding 10 m s^{-1} and air temperatures dropping from 0 to –31°C. The second storm was intermittent with wind speeds just exceeding 7 m s^{-1} and air temperatures between –20 and –32°C. Snowfall occurred and atmospheric water vapour remained unsaturated during both events. The latent heat flux measured by the eddy correlation system and calculated using the blowing snow model and the Dalton-type bulk transfer scheme are shown in Fig. 2(b). During the first blowing snow event, measured sublimation fluxes were large (30–60 W m^{-2}) and are strongly under-predicted by the bulk transfer calculation (typically 10 W m^{-2}) but better-predicted by the blowing snow model (30–55 W m^{-2}). During the second blowing snow event, measured sublimation fluxes were smaller (<10 W m^{-2}) and are modelled equally well by the bulk transfer or the blowing snow model. Over the periods shown, the blowing snow model (r^2 = 0.82, mean error =

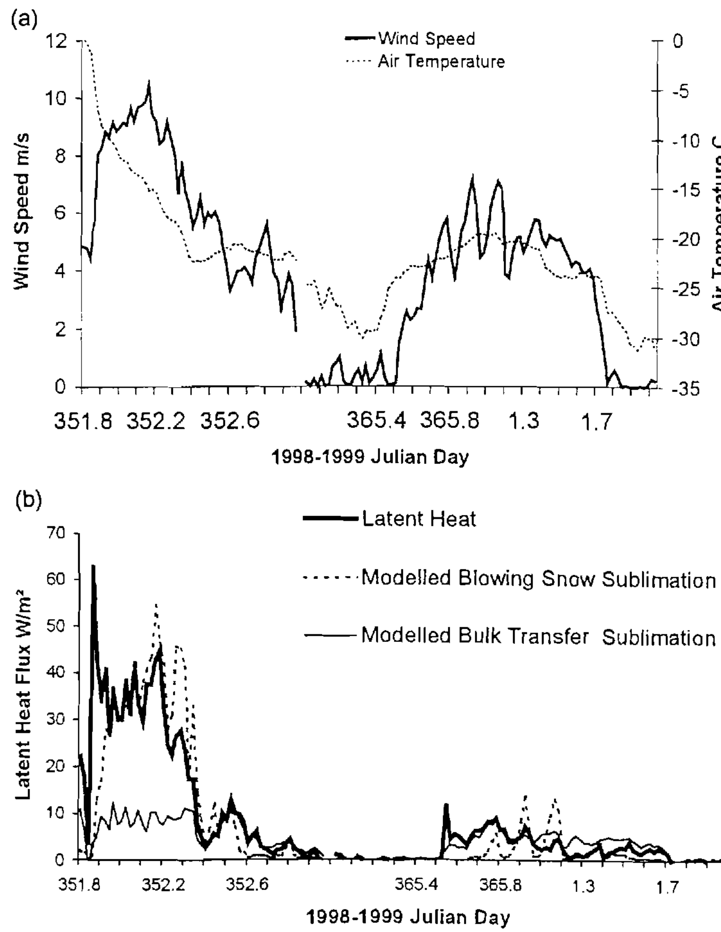

Fig. 2 Measurements at Kernen Farm near Saskatoon, 19–21 December, 31 December 1998 and 1 January 1999. (a) Wind speed and air temperature at 2-m height; (b) measured latent heat (sublimation) flux, sublimation calculated using a blowing snow model, sublimation calculated using a standard bulk transfer model.

-0.83 W m^{-2}, standard deviation of error = 7.6 W m^{-2}) performed better than the bulk transfer model (r^2 = 0.69, mean error = -3.7 W m^{-2}, standard deviation of error = 10.4 W m^{-2})

Turbulent fluxes during melt

A comparison of different Dalton-type bulk transfer turbulent flux schemes to measurements during stable conditions over melting snow was made for an isothermal,

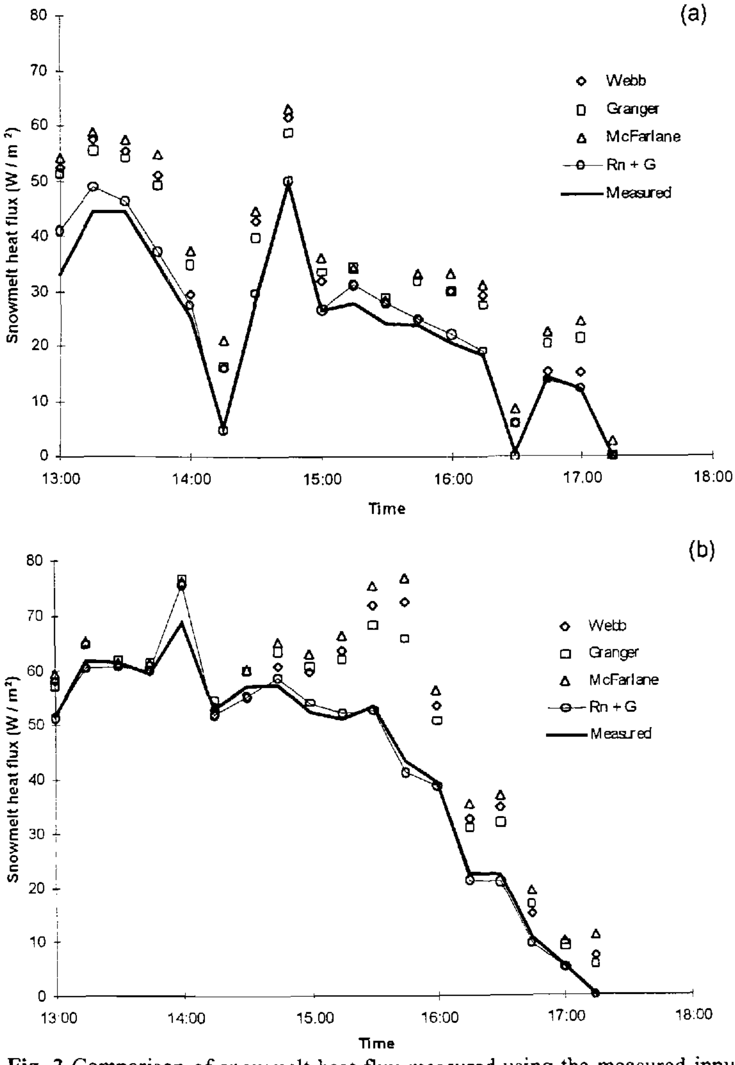

Fig. 3 Comparison of snowmelt heat flux measured using the measured inputs to the energy balance equation, and modelled using ground and radiation fluxes and estimated sensible and latent heat fluxes from the following schemes: Webb (1970), Granger & Male (1981) and McFarlane *et al.* (1992). The net radiation less ground heat flux is shown for further evaluation of the actual contribution from turbulent fluxes. (a) 11 March 1996; (b) 12 March 1996.

continuous snowpack near Saskatoon 11 and 12 March 1996 and is shown in Fig. 3. Three schemes were tested: the "log-linear" form commonly used for stable conditions (Webb, 1970), a modification developed by Granger & Male (1978) from measurements over a prairie snow cover, and a Richardson number formulation typical of the type used in large-scale atmospheric models (McFarlane *et al.*, 1992). All schemes were driven with an assumed surface temperature of 0°C and measured air temperature and humidity at 2 m height along with wind speed. For comparison, net radiation less ground heat flux is shown. *All* of the turbulent transfer schemes overestimate the downward (largely sensible) convective energy available for melt, the degree of overestimation depending upon the stability correction employed by the scheme. RMS errors in estimates of snowmelt heat fluxes over the 13:00–17:00 h period are shown in Table 1 for both days. Interestingly, the "best" melt rate simulation is obtained by disabling the turbulent transfer schemes and simply using net radiation and ground heat flux to estimate snowmelt.

Table 1 RMS errors (W m^{-2}) during isothermal melt of a continuous open snow cover in estimates of snowmelt heat fluxes by various commonly used turbulent exchange algorithms, compared to net radiation less ground heat flux. Kernen Farm, Saskatoon, 1996.

	Webb	Granger	McFarlane	$Q_n + Q_g$
11 March	10.0	10.0	13.0	2.8
12 March	10.7	9.0	13.0	2.0

Energy balance during snowmelt

An appreciation of the difficulties of modelling snow melt with even a relatively sophisticated land surface scheme is shown in the following section, with respect to the Canadian Land Surface Scheme, CLASS (Verseghy *et al.*, 1993). CLASS was chosen because it performs an energy balance on a distinct snow layer, contains standard bulk turbulent transfer formulae and a multi-layer soil model and has focused its performance for cold regions environments

Energy flux measurements collected near Saskatoon during 12 March 1996 were used to examine the performance of CLASS simulations of the melt of shallow, continuous, open environment snow covers. CLASS was run on an hourly basis using net longwave and incoming shortwave radiation, air temperature, wind speed and dew point. Initial snow and soil conditions were determined by field measurements. Figure 4 plots modelled and measured 30 min fluxes of energy fluxes during melt. There is poor association between modelled and "measured" latent energy (Fig. 4(a)) with an r^2 = 0.74, mean error = 2.48 W m^{-2} and standard deviation of error = 6.94 W m^{-2}. The largest relative deviations (10–15 W m^{-2}) are associated with the largest latent heat fluxes (positive or negative). In other words there is a general trend for CLASS to overestimate the measured values slightly with the degree of overestimation increasing with magnitude of latent heat. Figure 4(b) shows wide variations among "modelled" and "measured" sensible energy. Deviations are largest when measured sensible heat is nearly 0, but modelled sensible heat range from +2 to –28 W m^{-2}. The correlation between modelled and measured sensible heat has an r^2 =0.27, mean error = –4.5 W m^{-2} and standard deviation of error = 8.3 W m^{-2}. This poor correlation in sensible heat flux is in a large part due to errors in the estimate of snow surface temperature by CLASS. Figure 4(c) shows that the CLASS estimates of the ground heat flux are consistently much higher than the corresponding measured value. The correlation between modelled and measured values of ground heat flux was $r^2 = 0.34$, mean error = 11.4 W m^{-2} and standard deviation of error = 14.9 W m^{-2}. It is felt that the overestimation of ground heat flux is due to neglecting the influence of infiltration into unsaturated frozen soils on temperature profiles.

Figure 5 compares the 30 min melt fluxes calculated by CLASS with those calculated using net measured components in the energy equation (equation 2) on 12 March 1996 near Saskatoon. CLASS tends to underestimate the net energy available

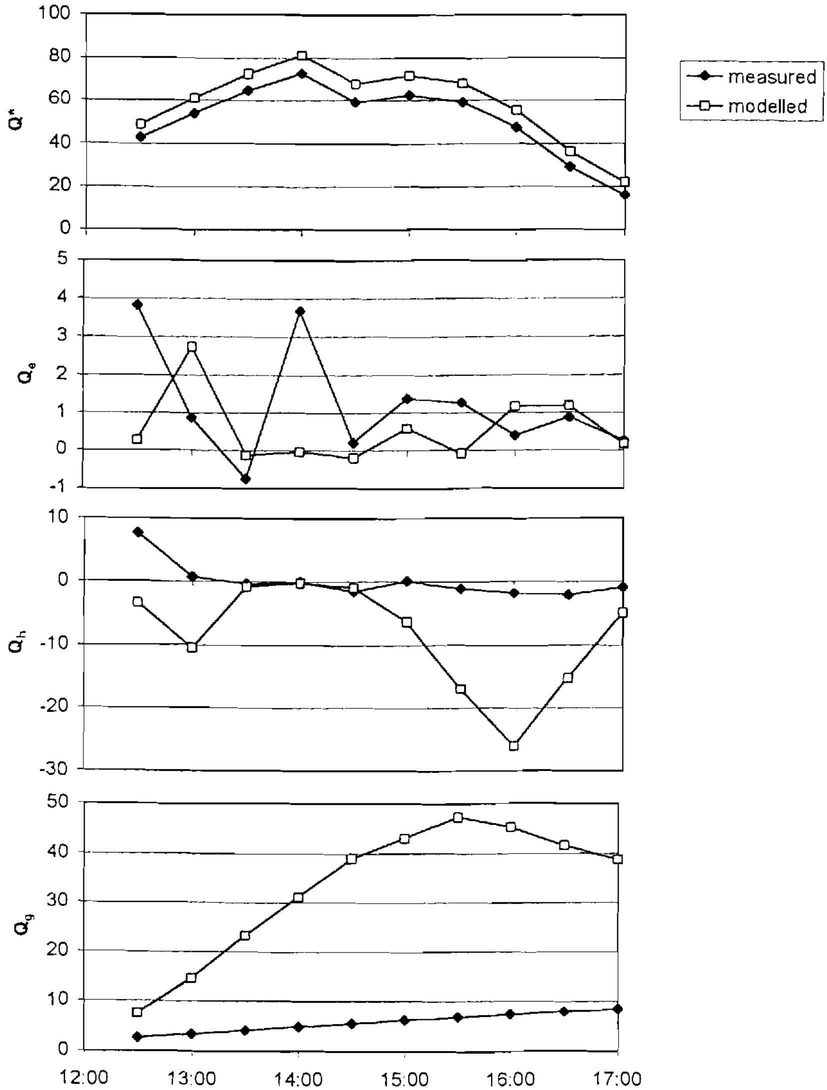

Fig. 4 Comparisons of modelled and measured 30 min fluxes at Kernen Farm, Saskatoon during melt, 12 March 1996: (a) net radiation, (b) latent energy, (c) sensible energy, and (d) ground heat.

for melt, with values ranging from 50 to 64 W m^{-2} during periods in the afternoon, and averaging 35 W m^{-2} over the interval 12:30–17:00 h. Another important feature demonstrated in Fig. 5 is the delay in timing of melt by CLASS. CLASS does not show melt occurring until 15:00 h. The overestimation of the ground heat loss by CLASS contributed 78% of the shortfall in the average melt rate flux; for the three day period it averaged 50% of the average melt rate flux. Such large ground heat fluxes due to heat conduction from the surface during snowmelt infiltration into frozen mineral soils are inconsistent with the small values found by coupled heat and mass transfer theory and measurement (Zhao *et al.*, 1997; Zhao & Gray, 1997).

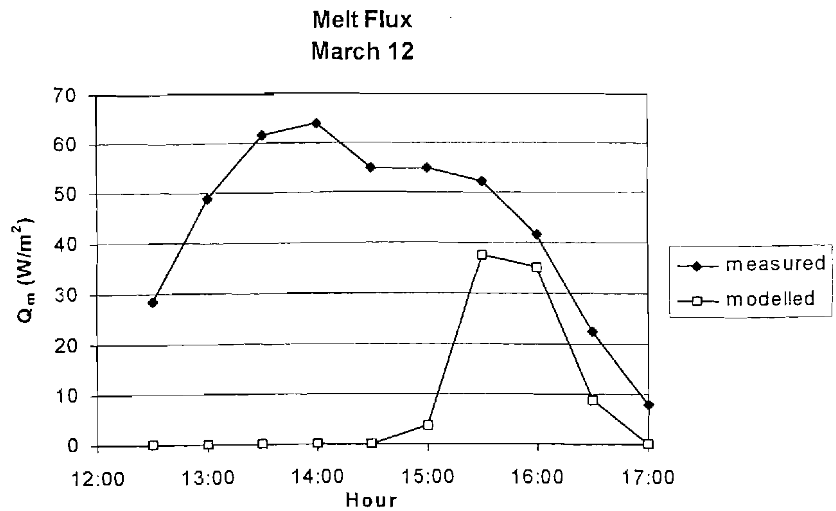

Fig. 5 Comparison of 30 min melt fluxes calculated by CLASS and corresponding values determined using measured components in the energy equation, Kernen Farm, Saskatoon, 12 March 1996.

IMPLICATIONS AND CONCLUSIONS

The following conclusions are made with respect to the next phase of snow process improvements in land surface schemes.

Land surface models do not presently characterize blowing snow redistribution and sublimation. These processes can cause substantial differences between accumulated snowfall and snow accumulation in prairie and arctic regions. Blowing snow sublimation losses are necessary to predict mean snow water equivalent at the grid-scale, whilst blowing snow transport should be calculated to gain an appreciation of the sub-grid variability of snow water equivalent. Latent heat fluxes during blowing snow can exceed 60 W m^{-2} for even a moderately-windy storm; these fluxes are poorly predicted by commonly-employed bulk turbulent transfer schemes but can be simulated using a blowing snow model.

Under stable conditions, melt of isothermal, continuous open environment snow covers is largely driven by radiative fluxes. Turbulent transfer schemes in land surface models calculate a turbulent contribution to melt that is larger than necessary.

Rapid and dramatic alterations in temperature gradient during snowmelt are caused by the downward conduction of latent heat released by the refreezing of percolating meltwater in cold snowpacks and frozen soils. Therefore, the processes of heat and mass transfers into frozen soils during infiltration can only be described properly by a multi-layered soil and snow model having a reasonably-small grid spacing. For those land surface models with only a few soil layers, the ground heat flux during infiltration into frozen soils should not be calculated by estimating the temperature gradient; a fixed, small value would provide a better result.

The CLASS land surface scheme underestimated the timing and rate of snowmelt in open environments because it grossly overestimated the ground heat loss during meltwater percolation and infiltration to frozen soils. Errors in the other energy terms for melt were compensatory and partly due to errors introduced to the snowpack energetics from ground heat.

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