Snow Modelling

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1. Snow, weather and climate

Snow in the Southern Hemisphere is largely confined to the permanent snowcover of the Antarctic ice sheet, but a large fraction of Northern Hemisphere land experiences seasonal snowcover. At the height of the Northern Hemisphere winter, snow covers approximately a third of the global land surface area. Snow is thus the most rapidly varying of surface covers, and it both influences and is influenced by climate far beyond the polar regions. Snow on the ground reflects a large fraction of incident solar radiation and absorbs large amounts of energy while melting, so it strongly influences energy exchanges between the land surface and the atmosphere, modifying overlying air masses (Ellis and Leathers 1999) and cyclone development (Ross and Wlash 1986, Elguindi et al. 2005). This impacts the climates of snow-covered regions locally (Cohen and Rind 1991) and remote regions through teleconnections (Gutzler and Rosen 1992, Cohen and Entekhabi, 2001).

Because snowcover influences the lower-boundary forcing of the atmosphere and varies slowly on the timescales of weather systems, there has been long-standing interest in snow for seasonal prediction. Blanford (1884) suggested a relationship between the strength of the Indian monsoon and Himalayan snowfall in the preceding season, and there have been many subsequent observational and modelling studies of connections between snowcover and monsoon circulations for both Eurasia and North America (e.g. Hahn and Shukla 1976, Barnett et al. 1989, Ferranti and Molteni 1999, Ellis and Hawkins 2001). Modulation of the North Atlantic Oscillation by snowcover has also been investigated (Saito and Cohen 2003, Saunders et al. 2003).

Decreasing snowcover and a consequent decrease in the surface albedo are expected to reinforce climate warming and contribute to the polar amplification of warming (Bony et al. 2006), but the predicted strength of this feedback varies widely between different global climate models (Cess et al. 1991, Qu and Hall 2006). Snow feedbacks are complicated by atmospheric responses, such as associated changes in cloud cover, and warming could lead to increased snow at high latitudes where it is limited by moisture supply rather than by temperature.

2. Snow models

Because snowcover strongly modifies surface characteristics and exchanges of heat and moisture between the surface and the atmosphere, representations of snow have to be included in land surface models for numerical weather prediction and climate modelling. Considerable efforts have been made in the development and evaluation of snow models for these and other applications. Operational hydrology requires efficient models that can be run with limited data for possibly poorly instrumented catchments; temperature index models have often been used for such applications (Ohmura 2001), but they require calibration and do not provide simulations of surface energy fluxes. A few sophisticated snow physics models have been developed for applications such as avalanche forecasting and remote sensing inversion that require detailed predictions of internal temperature, moisture and grain structure profiles in snow (Brun et al. 1989, Jordan 1991, Bartelt and Lehning 2002). These models are computationally expensive, require large numbers of parameters and are typically applied at point scales, rather on the large grid scales of atmospheric models. Snow models for NWP and climate modelling have generally been of intermediate complexity.

Snow models have to calculate the energy and mass balances of snow on the ground, shown schematically in Figure 1. The snow may be partially overlain by a vegetation canopy, which may itself hold intercepted snow, and the surface may be partially snow-free. Incoming solar radiation is absorbed and reflected according to the surface or canopy albedo, and incoming longwave radiation is absorbed and emitted according to the surface or canopy temperature and emissivity. Radiative fluxes beneath canopies are modified by interception of shortwave radiation and emission of longwave radiation by the canopy. Snow or rain falling on a surface of a different temperature will advect heat, but this is often neglected in models. Heat can be conducted or advected between snow and the ground at the base of the snowpack. Net radiation is partitioned into internal energy changes and turbulent fluxes of sensible and latent from the canopy, snow and snow-free ground. Snow and rain can be partially intercepted by vegetation canopies and subsequently removed by evaporation or sublimation to the atmosphere and unloading or drip to the underlying surface. Rain or meltwater at the snow surface can percolate into the snow, where a certain amount of water can be held in the liquid form or refreeze, releasing latent heat. Melt water reaching the base of the snow is partitioned into infiltration into the soil, runoff or basal ice formation.

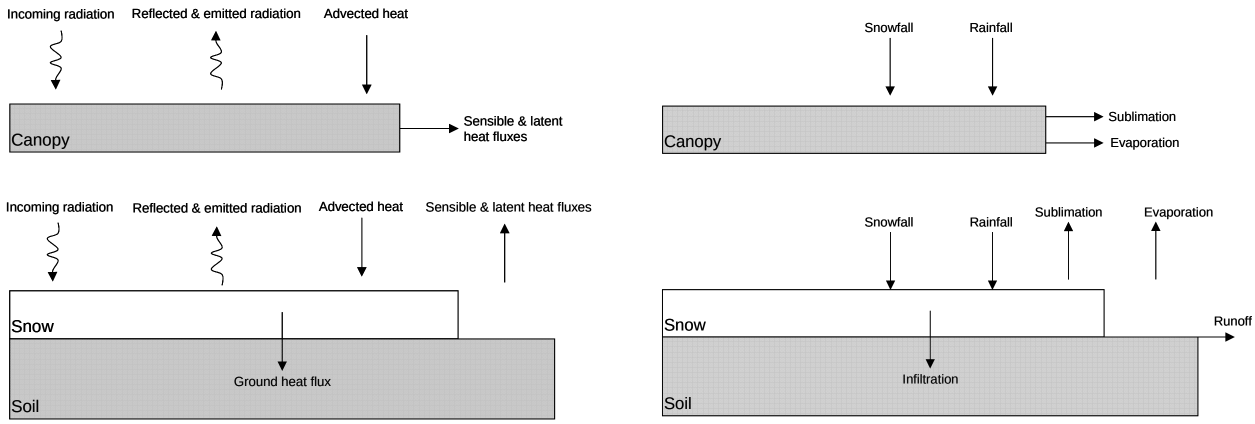

Figure 1. Schematic surface energy and mass balances.

Energy and mass balances are expressed through prognostic equations for temperatures and water contents of the canopy, snowpack and soil, coupled through evaporation, sublimation and melt terms. Simple models solve these equations for a single layer representing the snowpack or a combined snow and soil layer, but there has been increasing use of multi-layer snow models with 3 to 5 snow layers (e.g. Lynch-Stieglitz 1994, Oleson et al. 2004).

3. Parametrization of surface properties and processes

Albedo is an important parameter controlling the energy balance of snow. Snow albedos are typically parametrized as functions of surface temperature or age. Snow has a high albedo at visible wavelengths but a low albedo due to absorption by ice at near-infrared wavelengths. The mechanisms by which albedo changes with snow age also differ at different wavelengths: near-infrared albedos decrease as snow grain sizes increase during metamorphosis, whereas visible albedos decrease as the snow surface becomes contaminated by deposition of aerosols or vegetation litter. For these reasons, some models calculate albedos in two or more spectral bands. Several models use parametrizations of the form

$$\alpha(t+\delta t) = \alpha_{\min} + [\alpha(t) - \alpha_{\min}]\exp(-\delta t/\tau), \tag{1}$$

where $\alpha_{\min}$ is the minimum albedo for aged snow and τ is a time constant; one or other of these may be given different values for cold and melting snow. An albedo for fresh snow and an amount of snowfall required to refresh the albedo also have to be specified. Figure 2 compares snow albedos measured at Col de Porte in the French Alps with this parametrization using parameters from CLASS (Verseghy 1991); the broad pattern of albedo increasing when fresh snow falls and decreasing between snowfall events is captured. More sophisticated albedo schemes including grain size and soot content as parameters have been based on radiative transfer models of snow (Wiscombe and Warren 1980, Warren and Wiscombe 1980). The thermal emissivity of snow is invariably set to a constant value equal to or close to one.

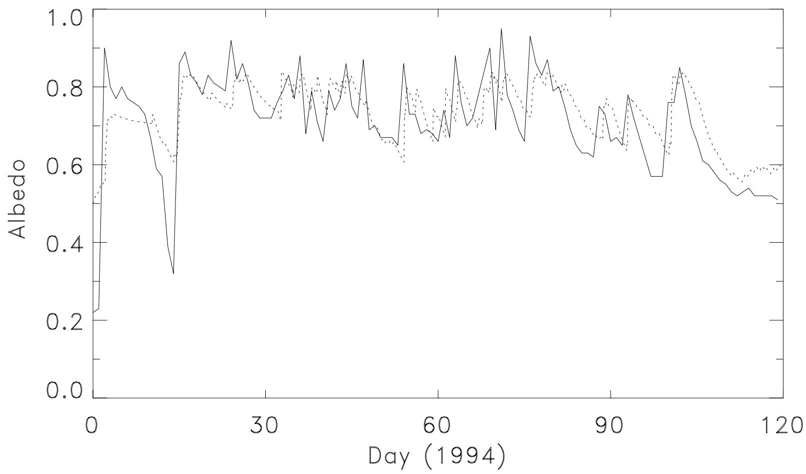

Figure 2. Observed (—) and simulated (· · ·) snow albedos.

Snow reduces the roughness of the surface by submerging vegetation and drifting into topographic depressions. This can be represented in models by decreasing the surface roughness length as a function of snow depth, down to a minimum value for deep continuous snow.

Gridboxes in large-scale models will often have partial snowcover, particularly in mountainous regions, or snow that is partially masked by vegetation. This has large influences on gridbox-average albedos and fluxes. Models represent this by parametrizing the snow-covered fraction as a function of gridbox-average snow mass, density and surface roughness, but differ widely in the fraction of snowcover that they predict for a given snow depth. Functions used have included

$$f_s = \frac{S}{S + 5\rho z_0} \tag{2}$$

(Douville et al. 1995),

$$f_s = \min\left[\frac{10S}{\rho}, 1\right] \tag{3}$$

(Verseghy, 1991) and

$$f_s = 0.95 \tanh(0.1S) \left(\frac{S}{S + 0.15\sigma_z} \right)^{1/2} \tag{4}$$

(Roesch et al., 2001), where S is the snow mass (kg m^{-2}), ρ is the snow density (kg m^{-3}), z_0 is the surface roughness length (m) and σ_z is the subgrid standard deviation of topography (m). Figure 3 shows snowcover fractions predicted by these functions for $\rho = 300$ kg m^{-3}, $z_0 = 0.02$ m and $\sigma_z = 0$ or 500 m. Some models use the snowcover fraction to calculate an average albedo for a gridbox or subgrid surface type, whereas others calculate separate energy balances for the snow-covered and snow-free fractions. This choice can have a large impact on the calculated rate of snowmelt when dominated by solar radiation (Liston 2004, Essery et al. 2005).

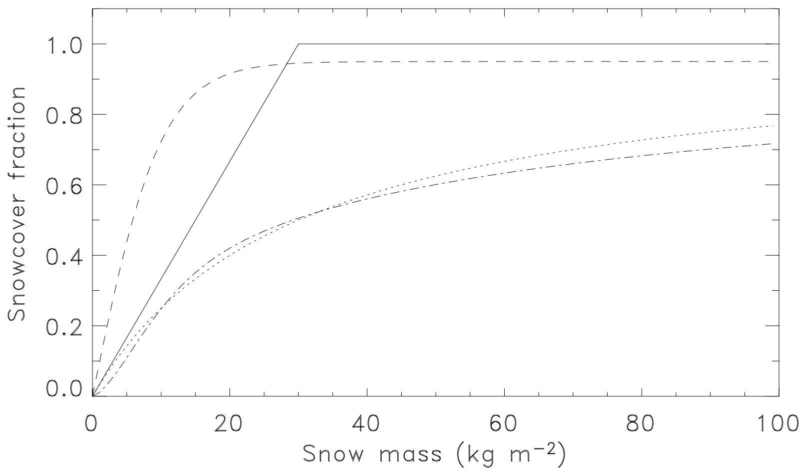

Figure 3. Snowcover fraction from Equations (2) (· · ·), (3) (—) and (4) with $\sigma_z = 0$ (- - -) or 500 m (- · ·-)

Sensible heat fluxes from the surface to the atmosphere are generally calculated using parametrizations of the form

$$H = \rho c_p C_H U (T_0 - T_a), \tag{5}$$

where ρ and c_p are the density and heat capacity of the air, C_H is an exchange coefficient, U is the windspeed, T_0 is the surface temperature and T_a is the air temperature. The exchange coefficient depends on surface roughness and atmospheric stability. There is particularly uncertainty in the form of exchange coefficients for stable conditions, which often occur over high latitude snowcovers. Monin-Obukhov similarity functions based on observations have a critical stability above which C_H is zero and the surface decouples from the atmosphere, potentially leading to unrealistically large radiative cooling, but the Louis (1979) formulation commonly used in large-scale models avoids this. The sensitivity of Antarctic climate simulations to the parametrization of surface fluxes has been investigated by King et al. (2001). Some snow models include a windless exchange coefficient to maintain sensible heat fluxes even at very low windspeeds (Jordan et al. 1999, Essery and Etchevers 2004).

Fresh snow can be readily eroded from the surface by wind, then trapped by vegetation or in the lee of topographic obstacles. Transport between grid boxes in a large-scale model will be negligible, but subgrid heterogeneities in the snow distribution are generated. Suspended snow has a large exposed surface area and can be subject to high sublimation rates, impacting the surface mass balance (Liston and Sturm 2004). Several specific models of blowing snow have been developed (e.g. Pomeroy et al. 1993, Gauer 1998, Xiao et al. 2000). These processes are not currently represented in global models but have occasionally been included in regional atmospheric models; applications have included simulating blizzards in the Arctic (Déry and Yau 2001) and surface mass balance in the Antarctic (van Lipzig et al. 2004).

4. Parametrization of internal snow processes

As snow ages, it increases in density due to grain metamorphosis and settling. The simplest models assign a constant value for snow density, but more sophisticated schemes allow for increases in snow density at a rate that may depend on snow temperature and mass. After snowfall, density is recalculated as a weighted average of the densities of new and old snow; fresh snow density may be assigned a constant value or parametrized as a function of air temperature and windspeed. A simple parametrization, similar to that given in Equation (1) for albedo, is

$$\rho(t+\delta t) = \rho_{\max} + [\rho(t) - \rho_{\max}]\exp(-\delta t/\tau), \tag{6}$$

where $\rho_{\max}$ is an upper limit on the density. Predictions from Equation (6) with default parameters of $\rho_{\max}$ = 300 kg m^{-3} and τ = 100 h underestimate measured snow densities at Col de Porte, shown in Figure 4, but better results are obtained by increasing $\rho_{\max}$ and τ to 450 kg m^{-3} and 340 h, respectively.

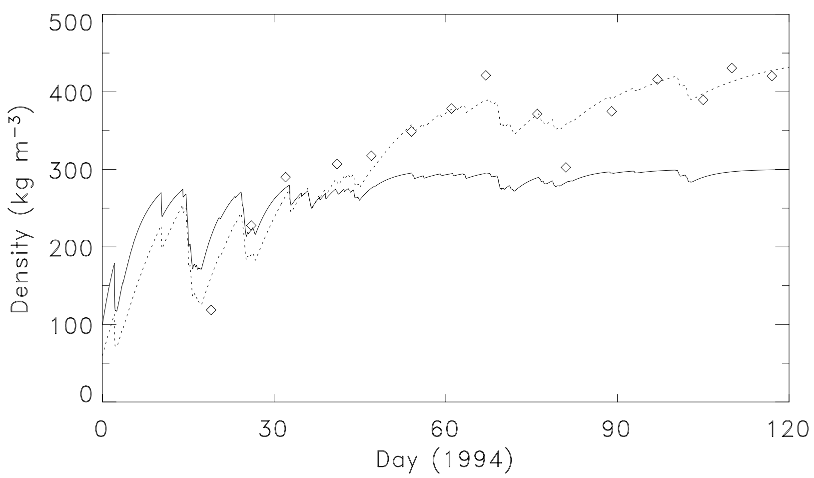

Figure 4. Observations of snow density (◊) and simulations with Equation (6) using default (—) and calibrated parameters (· · ·).

Snow has a low thermal conductivity, and this has important influences on soil temperatures and freezing under snowpacks. Transport of heat in snow is a complicated process involving conduction, advection, phase changes and radiation, but it is parametrized using effective thermal conductivities between layers of different temperature. Although strictly determined by snow microphysics, thermal conductivity is generally assigned a constant value or parametrized as an increasing function of snow density; parametrizations used in models have included

$$k = 0.074 + 2.576\times10^{-6}\rho^2, \tag{7}$$

$$k = 0.023 + 2.267(7.75\times10^{-5}\rho + 1.105\times10^{-6}\rho^2) \tag{8}$$

and

$$k = 2.22\left(\frac{\rho}{1000}\right)^{1.88}. \tag{9}$$

These are plotted in Figure 5 and compared with measurements from studies reviewed by Sturm et al. (1997).

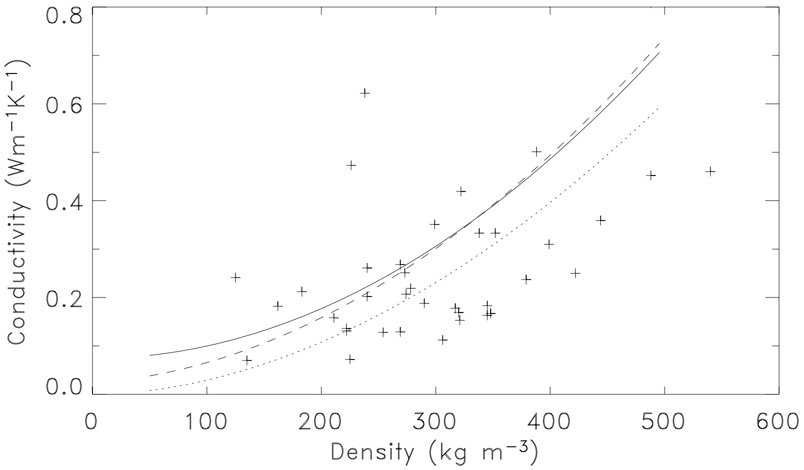

Figure 5 Thermal conductivity of snow from Equations (7) (—), (8) (- - -) and (9) (· · ·), and observations (+).

Although rarely represented in models, shortwave radiation can penetrate some distance into snow, causing subsurface warming in the snow or at the underlying surface (Koh and Jordan 1995). This can be parametrized by an exponential function with an extinction depth that depends on wavelength.

GCMs have generally neglected snow hydrology, removing any meltwater immediately from the snow to be partitioned into infiltration and runoff. A simple improvement allows for refreezing of liquid water in the snowpack, depending on the snow temperature. For snow at 0ºC a snow layer can be assigned a liquid holding capacity, which may depend on density, with excess water draining through the base of the layer.

5. Snow and vegetation

Exposed vegetation modifies the radiation, turbulent fluxes and precipitation at the underlying snow or ground surface. Snow falling on a canopy is partitioned into interception and throughfall; the interception capacity for snow can be much larger than for rain (Pomeroy et al. 1998), although some models take them to be equal. Intercepted snow can be removed from the canopy by sublimation, drip of meltwater or direct unloading. Several models use exponential unloading functions (Hedstrom and Pomeroy 1998) with timescales that may be related to air temperature and windspeed (Roesch et al. 2001).

There can be a large difference between the albedos of vegetated and open areas with snowcover; Viterbo and Betts (1999) found that reducing the maximum albedo from an unrealistically high value previously used over boreal forests greatly reduced a cold bias in ECMWF forecasts. Masking of snow albedo by exposed vegetation is represented in models by calculating a weighted surface albedo depending on snow depth and surface roughness, specifying a vegetation-dependent maximum surface albedo or using a radiative transfer scheme. Transmission of solar radiation is often parametrized using variants of Beer's law, giving the transmissivity for direct-beam radiation in the form

$$\tau = \exp\left(-\frac{G\Lambda}{\cos Z}\right), \tag{10}$$

where Z is the solar zenith angle, G is a projection coefficient depending on leaf orientation and Z, and Λ is an effective plant area index. Spectral values of G may be used to allow for the greater absorption by leaves at visible wavelengths, and the transmissivity may be integrated over Z for diffuse radiation. A more sophisticated two-stream radiative transfer scheme used by some models calculates absorption and multiple scattering of upwards and downwards radiation fluxes in canopies (Dickinson 1983).

Turbulent transfer through vegetation canopies is a complex process. Some models calculate separate fluxes from the canopy and the underlying ground (e.g. Sellers et al. 1986, Verseghy et al. 1993), but many models neglect this complexity.

6. Snow modelling for remote sensing and data assimilation

The high contrast in albedo with other surfaces makes it easy to map snowcover by remote sensing for areas that are not in darkness or obscured by clouds; NOAA has produced weekly, and now daily, maps of Northern Hemisphere snowcover extent using a succession of sensors since 1966 (Ramsay 1998). Indices based on reflectance in visible and near-infrared bands allow snow to be distinguished from clouds and allow the threshold for snow detection to be adjusted for areas with dense vegetation. Passive microwave data contain information on snow depth in addition to extent and may be obtained through clouds and under darkness. Simple algorithms based on differences in brightness temperature in different wavebands have often been used (Chang et al. 1987), but microwave emission from snowpacks is influenced by many factors, including temperature, grain size, water content and layer structure of the snow. There has therefore been interest in coupling radiative transfer models with snow models for inversion of remote sensing observations to obtain snow properties (Wiesmann et al. 2000).

Because snowcover strongly influences interactions between the land surface and the atmosphere, it is expected that assimilation of snow data can give improved skill in numerical weather forecasts. Manual and automated in situ snow depth measurements are widely available and have been used for data assimilation, but these observations are biased towards lower latitudes and lower elevations so may not produce results representative of large areas; remote sensing may address these problems (Drusch et al. 2004). The use of snow models with extended and ensemble Kalman filter techniques for data assimulation have been investigated (Sun et al. 2004, Slater and Clark 2006), the latter having the advantage that an adjoint model is not required. Optimal interpolation has been used both operationally and with a temperature index model to reconstruct gridded snow depths over North America (Brasnett 1999, Brown et al. 2003).

7. Uncoupled evaluation of snow models

There have been numerous evaluations of individual snow models run in uncoupled mode (i.e. driven with observed meteorological data rather than coupled to an atmospheric model) showing that they can simulate the accumulation and melt of snow well, sometimes after the introduction of model improvements (e.g. Lynch-Steiglitz 1994, Douville et al. 1995, Yang et al. 1997, Slater et al. 1998). Uncoupled model intercomparisons, of which there have been several for both local (e.g. Essery et al. 1999, Slater et al. 2001, Boone and Etchevers 2001, Etchevers et al. 2004) and regional (Bowling et al. 2003, Boone et al. 2004) scales, however, have found large spreads in model simulations, particularly while snow is melting and even for the comparatively simple case of uniform snowcover without projecting vegetation; an example from the Snow Model Intercomparison Project (SnowMIP) is shown in Figure 6. A further snow model intercomparison, SnowMIP2, is currently being undertaken to evaluate the ability of models to simulate snow processes under coniferous forest canopies and in forest clearings.

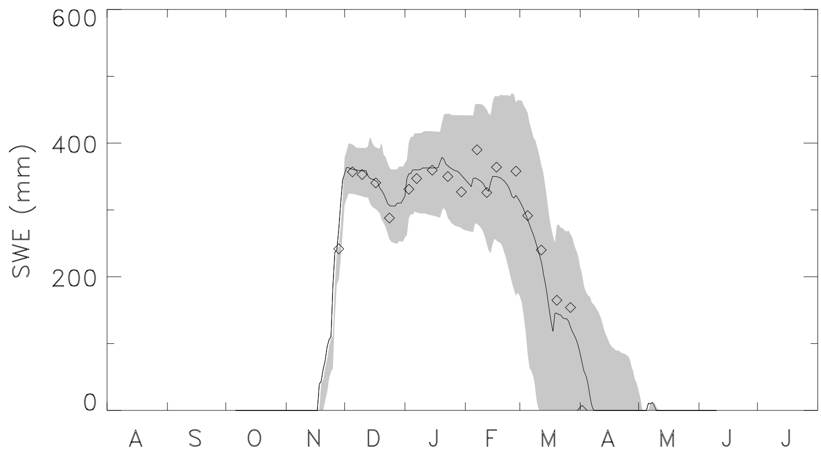

Figure 6. Observations (◊) of snow water equivalent at Col de Porte during the winter of 1996 - 1997, and a simulation with a calibrated model (—). The grey band shows the envelope of results from 22 uncalibrated models in SnowMIP.

Rather fewer studies have been published on the performance of models for snow on sea ice than on land (Ebert and Curry 1993). This is partly due to the greater difficulty of obtaining data, although observations have been used from Soviet drifting stations (Jordan et al. 1999) and the SHEBA experiment (Curry et al. 2001). Similar processes occur in snow on land or ice, but there are specific processes to be considered for ice, such as the loss of blowing snow to leads (Déry and Tremblay 2004), flooding by sea water (Andreas et al. 2004) and the influence of melt ponds on the surface energy balance (Fetterer and Untersteiner 1998).

8. Performance of snow models in climate and climate change simulations

A drawback of uncoupled simulations is that surface fluxes may be out of balance with changes in the driving meteorology, and models may not behave in the same way as they would when coupled to an atmospheric model and subject to feedbacks through the atmosphere. Snow models have also been evaluated in regional and global coupled simulations, both for individual models and in intercomparison projects. Gridded maps of snow extent from remote sensing and reconstructions are available for these evaluations.

Frei and Robinson (1998) compared simulations of snow extent by GCMs participating in AMIP with continental snow cover derived from visible satellite images. They found that the simulations underestimated snow extents in autumn and winter (especially for North America) and overestimated it in spring (especially for Eurasia). The interannual range in extent was underestimated, often by more than 50%. Frei et al. (2003) published a follow-up on snow in AMIP-2 simulations, finding the simulated seasonal cycle and interannual range of snow extent to be greatly improved over AMIP-1. Models still consistently overestimate snow extent over eastern Eurasia and underestimate it over western Eurasia. Interannual range is still underestimated, particularly for transition seasons. Frei et al. (2003) concluded that there is room to improve simulations of climate over continental Asia and parametrizations of precipitation and sublimation in cold, dry high elevation regions. In comparison with a gridded reconstruction of SWE for North America, variability between the AMIP-2 models was found to be significant in comparison with the components of the continental water balance (Frei et al. 2005). For coupled atmosphere-ocean models participating in the Fourth Assessment Report (AR4) of the Intergovernmental Panel on Climate Change, Roesch (2006) found that most overestimate spring SWE for both North America and Eurasia, mainly due to excessive snowfall.

A reduction in North American snowcover extent over the twenty-first century is a robust feature of the AR4 simulations (Frei and Gong 2005), but the trend varies between models. The change in surface albedos as the snowcover extent changes with increasing temperature was also found to vary widely between the AR4 models (Qu and Hall 2006). To constrain the snow albedo feedback, Hall and Qu (2006) compared the change in albedo for a given change in temperature in simulations of the current seasonal cycle and climate change; these were found to be highly correlated in the AR4 models but generally outside the range estimated from observations of the seasonal cycle.

9. Conclusions

Snow models are essential components of the land-surface schemes used in atmospheric models for numerical weather prediction and climate modelling. The sophistication with which snow processes are represented in these models is increasing, but there remain large differences in snowmelt and fluxes calculated by models in both uncoupled and coupled simulations. The increasing availability of ground-based and remote-sensing datasets will provide new opportunities for testing parametrizations of snow processes in a range of environments; evaluation of snow models for the scales on which they are typically applied in atmospheric models will be particularly valuable.

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