Modelling the effects of shrub-tundra on snow and runoff

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A thesis submitted for the degree of Doctor of Philosophy

University of Edinburgh
Centre for Ecology and Hydrology

2010
Abstract

Observational and modelling studies show that the warming of the Arctic is leading to shrub expansion. This shift in vegetation cover is expected to significantly alter the distribution of snow across the landscape and the interactions between the land surface and the atmosphere. Shrubs capture wind-blown snow, increasing snow depth and decreasing winter water loss through sublimation, and bend beneath the weight of snow, affecting albedo. Snow is highly insulative and affects the soil hydrological and thermal properties. Therefore, as the snow-vegetation-soil interactions is expected to be at the core of feedback loops leading to further shrub expansion, there is a need for models to be able to simulate these processes accurately. Initially using the community land surface model JULES (Joint UK Land Environment Simulator) this study investigates the effects of shrub-tundra on snow and runoff. Alternative formulations of soil processes are proposed, which are better adapted to the representation of subgrid heterogeneity in cold regions than the current model formulation, and evaluated over the Abisko and Torne-Kalix river basins. In addition, a high resolution shrub bending model, which calculates the exposed winter shrub fraction, is developed and parameterised for use alongside the snow cover parameterisation in JULES in order to provide a better representation of shrub-specific processes. This revised JULES more than doubles the efficiency coefficient and halves the negative bias between modelled and observed runoff in the shrub-tundra Abisko basin. However, the current structure of the model is found to be inadequate for use in investigating the effect of shrub-tundra expansion because it calculates a single energy balance for the snow-free and the snow-covered areas. To address this issue, a distributed three-source (snow-shrub-ground) model (D3SM) is developed. D3SM is evaluated against snow and energy flux measurements from a shrub-tundra basin in the Yukon, Canada, and is found to reproduce snowmelt energetics well. The effects of shrub expansion on the energy balance of the basin during snowmelt are then investigated by increasing the vegetation fraction and canopy height of the current shrub distribution, which is found to be positively correlated with topography. D3SM shows that the most significant effects of shrub expansion in the basin are to reduce the spatial variability of snow depth and to increase the sensible heat flux from the surface to the atmosphere.
Declaration

I declare that this thesis has been composed by myself and has not been submitted in any previous application for a degree. The work described is my own except where stated otherwise.

Cécile Bauduin-Ménard
July 2010
Acknowledgements

My sincere gratitude goes firstly to my supervisors, Dr Douglas Clark and Dr Richard Essery. Their patience, availability, guidance and support throughout the PhD has been invaluable. I am privileged to have worked with such professional and dedicated scientists at this early stage of my career.

I would also like to give a special thank you to Professor John Pomeroy for believing in me and for welcoming me within the IP3 network. It has been a joy and privilege to work and to argue with him.

I am indebted to the IP3 network who warmly accepted me into the Canadian hydrology community and provided me with much help and support over the past 3 years. A special thank you goes to Julie Friddel and to Jessica Boucher, Sean Carey, Qian Che, Ric Janowicz, Matt MacDonald and Mike Treberg for their help on the field.

I would also like to take this opportunity to thank the people who have helped me one way or another, often by simply taking the time to share their knowledge with me, especially Martin Best, Dan Bewley, Eleanor Blyth, Richard Ellis, Richard Harding, Andrew Hudson, Barbara Lachenbruch, Shawn MacDonald, Richard Milne, Karl Niklas, Erik Nilsen, Nick Rutter and Andy Wiltshire. Thank you also to my viva examiners, Ben Brock and Kate Heal, who made the experience far more enjoyable than I had dared to imagine.

This work would not have been possible without funding from the CLASSIC network, awarded by the Natural Environment Research Council, and without additional funds for field work and attendance at workshops by the IP3 network.

Last but not least, I would like to thank my partner, family and friends for supporting and encouraging me to return to academic studies after a 10-year break. A particular thank you goes to my Mum, Holly, Fiona and Michael for supporting me throughout these first years. My gratitude to Emma, on the other hand, is beyond words.
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Nomenclature

Acronyms

AS  Alpine site in the Wolf Creek Research Basin
BB  “Buckbrush” site in the Wolf Creek Research Basin
CL  Composite soil / snow layer scheme
D3SM Distributed 3-Source Model
DOY Day of the year
GB  Granger basin in the Wolf Creek Research Basin
GB2 Meteorological station in the valley in the Granger basin
JULES Joint UK Land Environment Simulator
LSM Land Surface Model
ML  Multiple snow layer scheme
NY  Frozen soil scheme derived from Niu and Yang (2006)
P1  Meteorological station on the plateau in the Granger basin
P2  Meteorological station on the plateau in the Granger basin
SBM Shrub Bending Model
TKB Torne and Kalix rivers basin
TRIP Total Runoff Integration Pathway
WIA Whitehorse International Airport
Roman alphabet

\[ a \]  Scale-dependent parameter in the \( F_{frz} \) parameterization (-)

\[ A_{ground} \]  Area of ground a shrub has to itself (m\(^{-2}\))

\[ A_{proj} \]  Cumulative area of primary branches (m\(^{2}\))

\[ b \]  Empirical soil-dependent constant (-)

\[ C \]  Dense canopy exchange coefficient (-)

\[ c \]  Time it takes for river flow to go through gridbox (s\(^{1}\))

\[ C_A \]  Volumetric heat capacity of the layer (J m\(^{-3}\) K\(^{-1}\))

\[ C_c \]  Areal canopy heat capacity (J m\(^{-2}\) K\(^{-1}\))

\[ C_e \]  Snow grain exposure coefficient (-)

\[ C_n \]  Areal heat capacity of the layer (J m\(^{-2}\) K\(^{-1}\))

\[ C_p \]  Heat capacity of dry air (J kg\(^{-1}\) K\(^{-1}\))

\[ C_{ice} \]  Heat capacity of ice (J kg\(^{-1}\) K\(^{-1}\))

\[ C_{water} \]  Heat capacity of water (J kg\(^{-1}\) K\(^{-1}\))

\[ d \]  Displacement height in D3SM (m)

\[ d \]  Distance across the gridbox in TRIP (m)

\[ d_h \]  Scaling depth for which SBM buries or reveals half of the shrub (m)

\[ d_r \]  Day of the summer solstice (day)

\[ D_w \]  Diffusivity of water vapour in air (m\(^{2}\) s\(^{-1}\))

\[ d_y \]  Average number of days in a year (day)

\[ d_{gb} \]  Distance across the gridbox (m)

\[ D_{in} \]  Inflow from neighbouring cells + cell runoff (kg s\(^{-1}\))

\[ D_{out} \]  Gridbox outflow (kg s\(^{-1}\))

\[ E \]  Moisture flux (kg m\(^{-2}\) s\(^{-1}\))
\( E \) Young’s or elastic modulus in SBM (N)

\( E_g \) Moisture flux over ground (kg m\(^{-2}\) s\(^{-1}\))

\( E_s \) Moisture flux over snow (kg m\(^{-2}\) s\(^{-1}\))

\( E_v \) Moisture flux over vegetation (kg m\(^{-2}\) s\(^{-1}\))

\( f \) Foliage factor in SBM (-)

\( f \) Saturated hydraulic conductivity decay factor in JULES-TOPM (-)

\( F(p) \) Complete elliptic integral of the first kind (-)

\( F_0 \) Gap fraction (-)

\( F_s \) Snow fraction (-)

\( F_v \) Fraction of shrubs protruding above the snowpack (-)

\( f_{eff} \) Effective foliage factor (-)

\( F_{frz} \) Impermeable frozen fraction (-)

\( F_{sat} \) Saturated fraction (-)

\( F_{v0} \) Snow-free vegetation fraction (-)

\( G \) Diffusive heat flux (W m\(^{-2}\))

\( g \) Gravitational acceleration (m s\(^{-2}\))

\( G_0 \) Ground heat flux (W m\(^{-2}\))

\( H \) Gridbox sensible heat flux (W m\(^{-2}\))

\( h \) Hour angle (rad)

\( h_c \) Canopy height (m)

\( H_g \) Sensible heat flux from the bare ground (W m\(^{-2}\))

\( H_s \) Sensible heat flux from the snow surface (W m\(^{-2}\))

\( H_v \) Sensible heat flux from the vegetation surface (W m\(^{-2}\))

\( I \) Intercepted snow mass in JULES (kg m\(^{-2}\))
\( I \) Second moment of area in SBM (m\(^4\))

\( I_0 \) Initial snow load on canopy in JULES (kg)

\( I_0 \) Solar constant (W m\(^{-2}\))

\( I_i \) Intercepted snow before unloading (kg m\(^{-2}\))

\( I_n \) Ice content in snow layer in ML (kg m\(^{-2}\))

\( I_{\text{max}} \) Maximum snow load on canopy in JULES (kg)

\( J \) Advective heat flux (W m\(^{-2}\))

\( K \) Hydraulic conductivity (kg m\(^{-2}\) s\(^{-1}\))

\( k \) von Karman constant (-)

\( k_t \) Atmospheric transmissivity (-)

\( K_{\text{sat}} \) Saturated hydraulic conductivity (kg m\(^{-2}\) s\(^{-1}\))

\( L \) Branch height (m)

\( L_c \) Latent heat of condensation (J kg\(^{-1}\))

\( L_f \) Latent heat of fusion (J kg\(^{-1}\))

\( L_s \) Latent heat of sublimation (J kg\(^{-1}\))

\( LE \) Gridbox latent heat flux (W m\(^{-2}\))

\( LW_g \) Net longwave radiation on the ground tile (W m\(^{-2}\))

\( LW_s \) Net longwave radiation on the snow tile (W m\(^{-2}\))

\( LW_v \) Net longwave radiation on the vegetation tile (W m\(^{-2}\))

\( LW_i \) Incoming longwave radiation (W m\(^{-2}\))

\( M \) Load applied on the branch in SBM (kg)

\( M \) Snowmelt heat flux in JULES and D3SM (W m\(^{-2}\))

\( m \) Shape defining parameter (-)

\( P \) Surface pressure (Pa)
$p$ \hspace{1em} \sin \alpha/2 \text{ in SBM (rad) }$

$PAI$ \hspace{1em} \text{Plant area index } (-)

$PAI_{eff}$ \hspace{1em} \text{Effective plant area index } (-)

$Q_a$ \hspace{1em} \text{Specific humidity at reference height (kg kg$^{-1}$)}

$Q_c$ \hspace{1em} \text{Canopy air space specific humidity (kg kg$^{-1}$)}

$q_l$ \hspace{1em} \text{Local downslope flow (m$^3$ s$^{-1}$)}

$q_s$ \hspace{1em} \text{Sublimation rate (kg m$^{-2}$ s$^{-1}$)}

$Q_{sat}$ \hspace{1em} \text{Saturation humidity (kg kg$^{-1}$)}

$R$ \hspace{1em} \text{Net radiation (W m$^{-2}$)}

$r$ \hspace{1em} \text{Branch radius in SBM (m)}

$r$ \hspace{1em} \text{Meandering ratio in TRIP (-)}

$r$ \hspace{1em} \text{Snow grain size (m)}

$R_c$ \hspace{1em} \text{Canopy net radiation (W m$^{-2}$)}

$R_r$ \hspace{1em} \text{Recharge rate (m s$^{-1}$)}

$r_{ag}$ \hspace{1em} \text{Bare ground aerodynamic resistance (s m$^{-1}$)}

$r_{as}$ \hspace{1em} \text{Snow aerodynamic resistance (s m$^{-1}$)}

$r_{av}$ \hspace{1em} \text{Aerodynamic resistance for vegetation (s m$^{-1}$)}

$r_a$ \hspace{1em} \text{Aerodynamic resistance (s m$^{-1}$)}

$r_i$ \hspace{1em} \text{Resistance from intercepted snow to canopy air space in JULES (s m$^{-1}$)}

$S$ \hspace{1em} \text{Intercepted snow mass in SBM (kg m$^{-2}$)}

$S$ \hspace{1em} \text{River water storage (kg)}

$s$ \hspace{1em} \text{Distance between 2 points on a slope in D3SM (m)}

$s$ \hspace{1em} \text{Length of a segment along the branch in SBM (m)}

$S_d$ \hspace{1em} \text{Snow depth (m)}
\( S_f \)  Snowfall rate \((\text{kg m}^{-2} \text{s}^{-1})\)
\( S_i \)  Incoming shortwave flux density normal to the beam \((\text{W m}^{-2})\)
\( S_M \)  Snowmelt rate \((\text{kg m}^{-2} \text{s}^{-1})\)
\( S_m \)  Snowmelt \((\text{kg m}^{-2})\)
\( S_n \)  Total soil moisture content of a layer \((\text{kg m}^{-2})\)
\( S_{\text{mass}} \)  Snow mass \((\text{kg m}^{-2})\)
\( S_{\text{dif}} \)  Diffuse solar radiation \((\text{W m}^{-2})\)
\( S_{\text{dir}} \)  Direct solar radiation \((\text{W m}^{-2})\)
\( S_{\text{max}} \)  Maximum canopy storage capacity \((\text{kg m}^{-2})\)
\( S_{\text{rel}} \)  Relative canopy storage capacity parameter (-)
\( S_h \)  Sherwood number (-)
\( SW_g \)  Net shortwave radiation on the ground tile \((\text{W m}^{-2})\)
\( SW_s \)  Net shortwave radiation on the snow tile \((\text{W m}^{-2})\)
\( SW_v \)  Net shortwave radiation on the vegetation tile \((\text{W m}^{-2})\)
\( SW_{\downarrow} \)  Incoming shortwave radiation \((\text{W m}^2)\)
\( SWE \)  Snow water equivalent (mm)
\( t \)  Time \((\text{s or hr})\)
\( T(z_{\text{wl}}) \)  Local transmissivity gradient \((\text{kg m}^3 \text{s}^{-1})\)
\( T_s \)  Potential surface skin temperature \((\text{K})\)
\( T_c \)  Canopy air space temperature \((\text{K})\)
\( T_g \)  Ground temperature \((\text{K})\)
\( T_m \)  Melting point \((\text{K})\)
\( T_s \)  Snow temperature \((\text{K})\)
\( T_v \)  Vegetation temperature \((\text{K})\)
\( T_{sl} \)  
Snow layer temperature (K)

\( T_{ss} \)  
Soil surface temperature (K)

\( U \)  
Snow unloading coefficient in JULES (-)

\( U \)  
Wind speed measured at \( z_u \) (m s\(^{-1}\))

\( u \)  
Effective river flow speed in TRIP (m s\(^{-1}\))

\( u \)  
Wind speed in D3SM (m s\(^{-1}\))

\( u_s \)  
Friction velocity (m s\(^{-1}\))

\( W \)  
Vertical flow of water in an unsaturated soil (kg m\(^{-2}\) s\(^{-1}\))

\( W_f \)  
Vertical flow of water in a frozen soil layer (kg m\(^{-2}\) s\(^{-1}\))

\( W_n \)  
Water content in snow layer in ML (kg m\(^{-2}\))

\( W_u \)  
Vertical flow of water in an unfrozen soil layer (kg m\(^{-2}\) s\(^{-1}\))

\( x \)  
Horizontal deflection of a branch in SBM (m)

\( Z \)  
Solar zenith angle (rad)

\( z \)  
Vertical deflection of a branch in SBM (m)

\( z_0 \)  
Effective momentum roughness length (m)

\( z_r \)  
Reference height in JULES (m)

\( z_t \)  
Temperature measurement height (m)

\( z_u \)  
Wind speed measurement height (m)

\( z_{0g} \)  
Ground tile roughness length (m)

\( z_{0s} \)  
Snow tile roughness length (m)

\( z_{0v} \)  
Vegetation tile roughness length (m)

\( z_{wd} \)  
Local water table depth (m)

\( z_{wmax} \)  
Maximum local water table depth (m)

\( \nabla q_t \)  
Horizontal divergence of snow transport
\( \overline{S} \)  Mean species-specific snow interception parameter (kg m\(^{-2}\))

\( \hat{S} \)  Solar radiation at a point (W m\(^{-2}\))

\( \Delta z_n \)  Soil or snow / soil composite layer thickness (m)

\( \overline{z}_w \)  Mean water table depth (m)

**Greek alphabet**

\( \alpha \)  Albedo (-)

\( \alpha \)  Deflection angle at the upper end of the branch in SBM (rad)

\( \alpha \)  Upslope area draining through a point in JULES-TOPM (m\(^2\))

\( \alpha_g \)  Bare ground albedo (-)

\( \alpha_s \)  Snow albedo (-)

\( \alpha_v \)  Vegetation albedo (-)

\( \alpha_{st} \)  Surface type albedo (-)

\( \alpha_{tile} \)  Tile albedo (-)

\( \beta \)  Slope angle (rad)

\( \chi \)  Topographic index (-)

\( \delta \)  Solar declination (rad)

\( \eta \)  Compactive viscosity (MPa s\(^{-1}\))

\( \iota \)  Empirical albedo decay parameter (-)

\( \lambda \)  Thermal conductivity of soil (W m\(^{-1}\) K\(^{-1}\))

\( \lambda_i \)  Thermal conductivity of ice (W m\(^{-1}\) K\(^{-1}\))

\( \lambda_s \)  Thermal conductivity of snow (W m\(^{-1}\) K\(^{-1}\))

\( \lambda_{CL} \)  Thermal conductivity of composite snow / soil layer (W m\(^{-1}\) K\(^{-1}\))

\( \Omega \)  Solar azimuth angle (rad)

\( \chi \)  Mean topographic index (-)
Φ  Latitude at the site (rad)
φ  Angle of a segment of branch to the vertical in SBM (rad)
Φ_T  Latitude at the tropic of Cancer (rad)
Ψ  Soil water suction (kg m\(^{-3}\))
ψ  Surface moisture factor (-)
ρ  Density of dry air (kg m\(^{-3}\))
ρ_i  Ice density (kg m\(^{-3}\))
ρ_s  Snow density (kg m\(^{-3}\))
ρ_w  Water density (kg m\(^{-3}\))
σ  Stefan-Boltzman constant (W m\(^{-2}\) K\(^{-4}\))
σ  Stress (Pa)
τ  Canopy transmissivity (-)
θ  Angle between the branch and the vertical in SBM (rad)
Θ_f  Unfrozen water content as a fraction of total soil moisture (-)
Θ_n  Total soil moisture content as a fraction of \(\theta_{sat}\) (-)
\(\theta_{sat}\)  Volumetric saturated soil moisture content (m\(^{-3}\) m\(^{-3}\))
\(\hat{\Omega}\)  Slope azimuth angle (rad)
ζ  Reducing factor in the calculation of \(G\) in CL (-)
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Chapter 1

Introduction
1.1 Introduction

Surface air temperature in the Arctic has increased faster than in any other part of the globe, at a rate of 0.4°C per decade since the mid-1960s. This warming has led to a decrease of approximately 10% of the areal coverage of snow, with a further 9 to 18% projected by the end of the century (Meehl et al., 2007). In addition, in spite of a 15 to 45% projected increase in snowfall, winter snowmelt events and a shortening of the snow season from both ends are expected to decrease the maximum snow depth over the Arctic, with the exception of Eastern Siberia and the islands in the Beaufort Sea (Figure 1.1) (Räisänen, 2008; Lawrence and Slater, 2010).

Paleoecological reconstructions and field experiments have shown that climatic change also affects Arctic vegetation, which responds quickly and dynamically to increases in temperatures, notably through expansion of deciduous shrubs (Payette et al., 1989; Brubacker, 1995; Chapin et al., 1995; Hobbie and Chapin, 1998). Increasing evidence from field observations (Figure 1.2), remotely sensed data and models suggests that the recent climate warming is indeed leading to a “greening” of the Arctic that can mainly be attributed to the encroachment of shrub-tundra on tundra landscapes (Sturm et al., 2005a; Jia et al., 2006; Raynolds et al., 2006; Tape et al., 2006). Here, tundra is defined as a land surface type where vegetation is less than 30 cm tall and shrub-tundra is defined as an ecosystem that occupies the latitudes and altitudes above the coniferous treeline, with “discontinuous and continuous canopies of deciduous shrubs of dwarf alder, willow and/or birch from roughly 30 cm to 3 m in height” (Pomeroy et al., 2006).

The relationship between shrubs and snow are at the core of feedback loops affecting the biochemistry, ecology, hydrology and energy balance of the Arctic (Figure 1.3). In the next sections, the factors that link the processes within the loop are described.
Figure 1.1: Projected changes (2080-2099 minus 1950-1969) in climatological snow-related properties. (a) Winter (November-March; NDJFM) snowfall in snow water equivalent. (b) Annual maximum snow depth. (c) Winter snow thermal conductivity. (d) Shoulder season (April, May, September, October; AMSO) snow cover fraction. (e) Day of year when autumn snow accumulation reaches 10 cm, grid cells with < 90-day or > 300-day snow season in either 1950-1969 or 2080-2099 are masked out in light grey (f) Day of year when spring snow melt brings snow depth below 10 cm, masking as in (e). Dark grey in all maps indicates gridboxes at least partly composed of glacier land cover type with perennial snow cover. Reproduced from Lawrence and Slater (2010).
1.2 Shrub - snow interactions

Shrubs increase the snow-holding capacity of the tundra by decreasing the near-ground wind speed downwind of and within shrub patches (Sturm et al., 2001a). Wind-blown snow is transported from open areas and stored in shrub-tundra landscapes.

The height, density and location of a shrub all affect its snow-holding capacity. Sturm et al. (2001a) reported that average snow depth in shrubby tussock tundra (mean canopy height = 0.13 m) was closer to snow depth in riparian shrubs (mean canopy height = 0.5) than snow depth in tussock tundra (mean canopy height = 0.1). They suggested that snow-holding capacity increased as a step-function of both the height and the density of the shrubs. However, Essery and Pomeroy (2004b) suggested that maximum snow-holding capacity is reached by approximately 1-m tall shrubs. Deep snow is also found in isolated patches or at the edge of large shrub-patches as they trap snow wind-blown from nearby open plains (Essery and Pomeroy, 2004b). On the other hand, snow within large shrub patches is shallower because it is only supplied by snowfall as surrounding shrubs prevent wind-blown snow transport.

By comparing modelled sublimation from wet- and moist-tundra and from shrub-tundra, Liston et al. (2002) found that sublimation was 68% smaller at the latter, where snow was 14% deeper than at the former. Although direct measurements of sublimation rates are difficult to obtain because of the volatile nature of snow particles, it is known to be the primary cause of winter water loss (Pomeroy and Gray, 1995). Benson (1982) used surface snow and snowfall measurements to estimate the redistribution of snow along the Arctic coast of Canada. He found that 58% of snowfall remained on the tundra until melt, 11% was transported to form drifts and 32% remained unaccounted for, suggesting that it had sublimated. This estimate was later corroborated by Pomeroy et al. (1998) who found that sublimation could account for up to 40% of water loss in boreal forests.

1.3 Snow - soil temperature interactions

The internal structure of a snowpack provides low thermal conductivity to the underlying surface, typically 0.08 and 0.42 W m\(^{-1}\) K\(^{-1}\) for fresh and old snow respectively (Oke, 1987). Snow particles are separated by air and motionless gas provides excellent insulation. Furthermore, for the same snow depth, the thermal conductivity of the snowpack is expected to be affected by a transition
Figure 1.2: Comparative photographs taken in 1948 and 2002, showing the increase in shrub abundance along the Colville River, northern Alaska. Dark objects are individual shrubs 1 to 2 meters high and several meters in diameter. Reproduced from Sturm et al. (2005a).

Figure 1.3: Schematic representation of feedback loops between processes at the surface and climate warming. Based on Sturm et al. (2005a) and Chapin et al. (2005).
from open to closed shrub-tundra because of a change in the internal structure of the snowpack (see Figure 1.1c). In a windswept open tundra environment, a snowpack offers little resistance to heat loss because it has been compacted and contains relatively highly conductive hard wind slabs that result from snow particles breaking into small ice grains (Liston et al., 2002). Below these slabs is a highly insulative, poorly bonded snow (depth-hoar) that forms in response to large temperature gradients. Shrub expansion will therefore prevent the formation of the top wind-swept layer and increase the percentage of depth hoar, thus decreasing the conductivity of the layer (Sturm et al., 2001a; Liston et al., 2002). For example, Sturm et al. (2001a) found that the cumulative thermal effect of shrubs over winter was a 2.7-fold difference in winter heat loss compared to sparse tussock tundra.

1.4 Soil temperature - carbon cycle interactions

Sturm et al. (2005a) observed a difference of 29°C (-45°C and -16°C respectively) between air and soil temperatures, measured at 0.3 m depth, under a 0.5 m deep snow pack. Sturm et al. (2001a) also found that soil temperature below shrubs dropped below -6°C, which is the thermal limit for microbial activity, on average 50 days later than under sparse tussock tundra. This suggests that the insulative properties of a snowpack could promote biotic productivity below ground even when above ground temperatures are below optimal temperature for plant or microbial activity. The warmer winter temperatures observed beneath shrubs could enhance nitrogen (N) mineralization by about 170 mg of N m⁻² year⁻¹, which could support an increase in plant production of about 15 g m⁻² year⁻¹ and, therefore, promote further shrub growth (Shaver and Chapin, 1991; Chapin et al., 2005; Buckeridge and Grogan, 2008). The net effect of enhanced microbial activity on the carbon cycle is uncertain. On the one hand, shrub expansion increases above ground biomass and, therefore, stimulates carbon fixing (Sturm et al., 2005a,b), acting as a carbon sink. On the other hand, enhanced winter microbial activity under deep snow promotes CO₂ respiration and acts as a source (Oechel et al., 1997; Fahnestock et al., 1998; Morgner et al., 2010).

Warmer soil temperatures also affect the carbon cycle by increasing the temperature of the active layer, the layer at the top of the permafrost subject to freezing and thawing on an annual basis. Because most exchanges of energy, moisture and gases between the atmosphere and the permafrost occur through this layer, changing its thermal regime could trigger the release of significant
amounts of greenhouse gases to the atmosphere (Fukuda, 1994; Michaelson et al., 1996; Anisimov et al., 1997). Of particular importance for the feedback on climate change are the large amounts of methane whose global warming potential over 100 years is 25 times greater than CO$_2$ (Forster et al., 2007).

1.5 Snow - soil - runoff interactions

Snowmelt contributes up to 80% of the annual runoff in the northern parts of central and eastern Siberia and to about 50% in northern Europe and northeastern Canada (Walsh et al., 2004). At high latitudes, runoff is characterised by low winter flows, a high spring snowmelt-induced event and a recession rainfall-induced period which lasts until the beginning of the snow season. Peterson et al. (2002) found that warmer air temperatures and increased precipitation have led, between 1936 and 1999, to a 7% increase in annual freshwater inflow from the large Arctic rivers to the Arctic Ocean. However, more recent observational records have shown a 9.8% increase from 1977 to 2007 (Overeem and Syvitski, 2010). Of most significance for the management of water resources is the change in the seasonal runoff characteristics. While mid-winter snowmelt events are increasing winter flow (e.g. Serreze et al., 2002; Walsh et al., 2004; Hinzman et al., 2005), the reduction of the snow season is causing peak flow to occur earlier. For example, Overeem and Syvitski (2010) noted a 66% increase in runoff in the month preceding peak flow (May) but a 7% reduction in peak discharge. In summer, warmer air temperature is expected to reduce runoff but increase evapotranspiration (see e.g. Serreze et al., 2002; Dankers and Christensen, 2005).

Shorter snow seasons will delay freeze-up, (i.e. when a body of water becomes covered by immobile ice), and cause river-ice break-up and nival floods to occur earlier in the season. Such changes will require adaptation strategies for water resources management in Northern communities as unusual break-up patterns could cause structural damage to buildings and communications systems, and different river level could affect transportation or hydro-electric energy infrastructures (NRTEE, 2009).

Changes to the frozen soil and permafrost regimes of the region are also likely to affect flow paths and flow rates. In zones of continuous permafrost (90-100% permafrost) the ice-rich layer acts as an impermeable layer where surface runoff leads to large snowmelt-induced peak flows which contribute little to groundwater recharge. The ratio of surface to sub-surface runoff decreases with decreasing permafrost distribution as more snowmelt water is able to infiltrate the soil column.
and contribute to the groundwater recharge (Woo, 1986). At the global scale, increases in the input of freshwater in the Arctic ocean could affect global climate (e.g. Broecker, 1997; Walsh et al., 2004). The Atlantic thermohaline circulation is driven by the temperature and the salinity of the water, both of which determine the density of the water. Abrupt climatic change in the last glaciation has been linked to a reduction in North Atlantic Deep Water formation, driven by the density of the surface water (Broecker, 1987; Rasmussen et al., 1996; Broecker, 1997). The relatively warm climate experienced in Northern Europe, compared to the much colder one experienced on the other side of the Atlantic at the same latitudes, has long been identified as being promoted by warm water from lower latitudes that flows up to the northeast of Iceland and undergoes evaporative cooling (e.g. Maury, 1855). Stocker and Schmittner (1997) and Manabe and Stouffer (1993) predicted a major slow down in the thermohaline circulation by increasing CO$_2$ levels 1% each year for 140 years, which led Broecker (1997) to conclude that increased fresh water inflow to the Arctic Ocean due to global warming could foster changes to the climate of Northern Europe similar to those experienced in the last glaciation. This hypothesis is however controversial and model experiments by Seager et al. (2002) suggested that the thermohaline circulation only had a minor effect on temperatures in Northern Europe.

### 1.6 Snow-energy balance interactions

The albedo of snow-free shrub-tundra can change from 0.15 to 0.9 following a single snowfall event, changing the radiation budget of a landscape on a very short timescale (Sturm et al., 2001a). Sturm et al. (2005b) calculated that, in winter, the difference in absorbed solar radiation between a shrub-free tundra and shrubland could be of the order of 69 to 75%, despite the limited sunlight available at high latitude from October to March. Investigations into snowmelt energetics over tall shrub-tundra sites and nearby dwarf shrub sites in the Yukon (Pomeroy et al., 2006; Bewley et al., 2010), in the Northwest Territories (Marsh et al., 2010) and at five sites in Alaska (Sturm et al., 2005b) showed that the presence of tall shrubs led to an increase in absorbed solar radiation caused by the lower albedo of the shrub branches protruding above the snowpack. This, in turn, led to the direction of energy fluxes being from the surface, because of the warm shrub branches, to the atmosphere, even if snow remained on the ground. Compared to wet- or moist-tundra, shrub-tundra also promotes the transfer of heat between the surface and the atmosphere because tall shrubs increase the roughness of the
surface and, by extension, the turbulent exchanges. This increase in absorbed radiation was estimated to have accounted for 2% of the recent warming caused by land-surface change (Chapin et al., 2005).

Tall shrubs provide shading, reducing the amount of shortwave radiation reaching the snowpack (Bewley, 2006; Bewley et al., 2010). However, the loss in shortwave radiation is compensated by longwave radiation and sensible heat flux from the canopy to the snow, such that net radiation is higher below shrubs than for exposed snow. Sturm et al. (2005b) and Pomeroy et al. (2006) observed melt rates to be generally higher at tall shrub sites than at dwarf shrub sites, which lead to both landscapes being snow-free at the same time, even if snow was initially deeper under tall shrubs. However, this general but not systematic pattern is a result of the non-linear relationship between snow depth and shrub height, or in other words, of shrub exposure in the winter. The effect of shrubs on winter albedo is moderated by shrubs bending and getting buried in the snowpack. As a consequence, all else being equal, energy exchanges and melt rates at a buried, bent shrub site are no different to those at a site without vegetation.

Sturm et al. (2005b) and Pomeroy et al. (2006) suggested that bending occurred as a result of a combination of snowpack properties, temperatures and wind. This behaviour is a common evolutionary trait amongst shrubs in cold environments. Johnson (1987) found the same adaptation mechanism in *Salix glauca*, a species of willow shrub that can be found at the top of avalanche-prone slopes in the Canadian Rockies. Following tests on lodgepole pine, Engelmann spruce, glandular birch and willows, he found that, although *S. glauca* was not the only plant to respond to loading by bending in order to avoid structural damage, it had the highest elasticity and could sustain the highest bending stress. Equally, Beismann et al. (2000) measured the brittleness of 8 species of willows in south west Germany and found that the least brittle species were those evolving in alpine environments. Beismann et al. (2000) suggested that the biomechanic properties of the species evolved in response to the geographical distribution of the shrubs where, to survive the winter conditions, they adapted to sustain snow loads and avalanches.

1.7 Modelling the interactions

The uncertainties related to the complex relationships between Arctic processes have stimulated efforts to develop better representations of high latitude processes in large-scale models.
Global predictions, presented in reports like the Intergovernmental Panel on Climate Change 4th Assessment Report (Solomon et al., 2007) or the Arctic Climate Impact Assessment (ACIA, 2005), are based on results from General Circulation Models (GCMs). GCMs comprise three components, an atmospheric, an ocean and a land surface model (LSM), which are able to investigate physical processes in the Earth system. Historically, LSMS evolved from being simple parameterization schemes intended to represent the boundary conditions on the land surface in GCMs, to stand-alone models used to simulate land surface processes related to, for example, the water or carbon cycles.

Model intercomparison projects like the Project for Intercomparison of Land-Surface Parameterization Schemes (PILPS) (Henderson-Sellers et al., 1995) have motivated improvements in LSMS. Two of the projects, PILPS phase 2(d) (Slater et al., 2001) and 2(e) (Bowling et al., 2003a), were designed to evaluate the treatment of cold region hydrology in uncoupled land-surface parameterizations. Of particular importance for this study was the investigation of shrub-tundra hydrological processes in PILPS 2(e).

These two intercomparison projects provided insights into the relationship between modelled high latitude processes. In PILPS 2(d), Luo et al. (2003) found that an accurate representation of snow processes was the most important feature in models for investigation into cold region hydrology because it

1. Determines water availability for runoff and infiltration.
2. Affects the albedo of the landscape and thus the net radiation at the surface.
3. Influences soil temperatures through its insulating properties.

As part of the same experiments, Slater et al. (2001) showed that the more structurally complex snow models were also able to provide better simulations of snow water equivalent than the simple models.

However, further investigations into cold regions processes in PILPS2(e) showed that small changes in simple parameterizations could also have significant effects on modelled processes. For example, Nijssen et al. (2003) identified the treatment of the snow cover fraction, related to albedo and thus net radiation, and of the roughness of the surface elements, effective in energy exchanges, as being critical to the representation of the energy balance. With regards to river flow, they found that the partitioning between surface and subsurface runoff and water storage determined the treatment of the snowmelt-induced runoff peak. Models that classified their runoff mostly as surface runoff were found to overestimate the
runoff peak and underestimate the recession period, while sub-surface dominated schemes were found to delay runoff.

The findings from PILPS and other cold region model intercomparison projects like SNOWMIP (Etchevers et al., 2002) and SNOWMIP2 (Essery et al., 2009), along with increasing computational efficiency, have allowed models to evolve and motivated a large number of improvements in LSMs. However, the more that is understood about land surface processes, the more the complexity of the land surface becomes apparent, necessitating ever more research.

1.8 Aim of thesis

The aim of this thesis is to contribute to the effort of improving the understanding of land surface processes by assessing the effects of shrub-tundra on snow and runoff. The objectives of this study are to:

1. Improve model representations of snow-shrub interactions.
2. Improve model representations of the hydrology in shrub-tundra environments.
3. Assess the effect that shrub expansion could have on the hydrometeorology of Arctic basins.

These objectives will be addressed in the subsequent chapters and in the order described below by investigating some of the interactions described in the previous sections and in Figure 1.3. In Chapter 2, a description of the sites and the data used in this study is presented. In Chapter 3, the physics of the land surface model used in this study, the Joint UK Land Environment Simulator (JULES), is given. In Chapter 4, alternative formulations of soil processes in JULES are introduced and evaluated against observed streamflow. The need for a better representation of shrub processes in JULES is identified and leads to the development of a shrub bending model (SBM) described and evaluated in Chapter 5. The parameterization derived from SBM for use in JULES is tested and evaluated in Chapter 6. The limitations of JULES with regards to the simulation of land surface processes in a heterogeneous landscape are addressed in Chapter 7, where a surface energy balance model calculating separate energy balances for each surface, the Distributed 3-Source Model (D3SM), is presented and evaluated against snow depth and turbulent fluxes. In Chapter 8, this model is used to assess the effect of shrub expansion on snowmelt energetics in a sub-Arctic basin. Finally, the work from the previous chapters is summarised in
Chapter 9.
Chapter 2

Description of the study areas and of the meteorological and evaluation datasets
2.1 Introduction

In order to attain the aim and objectives stated in Section 1.8, one sub-arctic and one arctic basin were selected to evaluate the performance of the models used in this study: the Wolf Creek research basin (WCRB), Yukon Territory, Canada and the Torne and Kalix river basin (TKB), Fennoscandia. The sites were chosen because they span a range of ecological zones representative of high latitude ecosystems and have been part of long-term monitoring campaigns. The data available for each basin are complementary and will enable the study of the effect of shrub-tundra over a range of scales and processes. Although the information available at the two sites can not be directly compared because of the difference in the resolution of the data (see the following sections for detail) and in the time period the data were collected, the meteorological data reflect the continental climate at the WCRB and the maritime climate at the TKB. For example, the data available shows that mean annual snowfall in the shrub-tundra basin in the TKB, Abisko, is 334 mm (1979-1998) against 190 mm and 127 mm for the two years of assimilated snow data in the WCRB (1998-1999 and 2003-2004). Equally, mean annual air temperature in the WCRB is 2.4°C lower (-4.1°C, April 2006 to April 2008) than in Abisko (-1.7°C, April 1979 to April 1998) although Abisko is 8° North of the WCRB.

Manual snow and vegetation surveys and eddy covariance data available over the WCRB will be used to study high resolution processes in a small shrub-tundra basin and to investigate potential changes in these processes associated with the predicted expansion of shrub cover. Data over the TKB will be used to study large scale hydrological processes and their representation in a complex land surface model. Work performed in the two basins is linked through the implementation of new parameterizations derived in one basin, for example the exposed vegetation cover, and used to inform the representation of processes in another. A more detailed description of the sites and data used in this study can be found in the following paragraphs.

2.2 Wolf Creek Research Basin

The WCRB is a 195 km² watershed situated 15 km south of Whitehorse at 60°35’N, 135°11’W in the southern headwaters of the Yukon river drainage basin (Bewley et al., 2007) (Figure 2.1). The basin lies between 640 and 2100 m.a.s.l. and thus provides characteristics of various Arctic ecosystems over a relatively
small area. Mean annual temperature at the Whitehorse International Airport (WIA, 706 m.a.s.l) is -0.7°C, with January being the coldest month (mean of -17.7°C) and August being the warmest (mean of 12.5°C) (1971 - 2000, Environment Canada, 2008). Mean annual precipitation at WIA is 267 mm, 39% of which falls as snow although Pomeroy et al. (1999) found that snowfall in WCRB exceeded that at the airport by 33% to 47%.

Although WCRB is conventionally classified as being in a zone of sporadic discontinuous permafrost (defined as 10-50% permafrost cover in Brown et al., 1988, and Heginbottom and Dubreuil, 1993), Lewkowicz and Ednie (2004) found that the entire range of permafrost conditions, which commonly extends across hundreds of kilometres in lowland permafrost areas, is present in the basin.

Meteorological and field data were collected at three sites within the WCRB, each covering different Arctic landscape characteristics: tall shrub (“Buck brush” site), shrub-tundra (Granger basin) and tundra (alpine site) (Figure 2.2). The three sites are described in the following sections.

Airborne LiDAR data over the WCRB were collected and processed by Chasmer et al. (2008), Hopkinson and Chasmer (2009) and Richard Essery, University of Edinburgh. The data provided all the digital elevation, canopy height and
fractional vegetation cover ($F_v$) maps for the WCRB, used in the figures and as land surface characteristics in Chapters 7 and 8. LiDAR data were obtained by sending a single laser pulse from the sensor (Airborne Laser Terrain Mapper 3100, Optech Inc., Toronto, Canada). The time delay between the transmission of the pulse and the detection of the backscatter element (or “return”) was then used to calculate the distance between the sensor and the surface. Information on the laser pulse ranges, aircraft movement and ground and airborne GPS trajectories, allows the returns to be classified into ground or near-ground returns and to give surface elevation and canopy height maps respectively. The vegetation fraction was then calculated as the ratio of the number of canopy laser returns to total returns (Chasmer et al., 2008).

Figure 2.2: Meteorological stations at the sub-basins within the Wolf Creek Research Basin: (a) P2 (b) GB2 (c) P1 (d) Alpine (e) BB.
2.2.1 The shrub-tundra site: Granger basin

The Granger basin is situated within the sub-alpine ecozone, characterised by a shrub-tundra landscape (Figure 2.3 and Figure 2.4). It is 8 km² and ranges in elevation from 1295 to 1555 m.a.s.l. The stream flows southeast along the valley which leads to the landscape being dominated by a north and a south facing slope (17° and 13° respectively) and a level valley bottom (Quinton et al., 2004).

At least 15 species of tall willow shrubs (on average 180 cm tall from field measurements), including Salix glauca and Salix pulchra, cover the riparian zones at the bottom of the valley (Dr Michael Treberg, Carleton University, personal communication). Some of these shrubs get buried under the snowpack in the winter and rapidly “spring up” during melt (Pomeroy et al., 2006; Bewley et al., 2007). Birch shrubs are also widespread within the basin. Their height ranges from 0.3 to 1.5 m. Their distribution is stratified with elevation, the shorter shrubs being generally on the exposed plateau while the taller shrubs are found near the valley bottom.

Data from three meteorological stations situated in GB are used in Chapter 7 and Chapter 8:

- “P2” located on an exposed plateau away from the drainage course at 1424 m.a.s.l., about 330 m away from GB2. Shrubs at P2 are birch and on average 0.32 m high (Bewley et al., 2010) (Figure 2.2-a).
- “GB2” located in the riparian zone amongst the willows at 1363 m.a.s.l. (Figure 2.2-b).
- “P1” located about 90 m west of P2 within a localised depression on the plateau (Figure 2.2-c). The site comprises both birch (< 1 m) and willows (< 2.5 m) which lead to more snow being trapped than at P2.

The air temperature, wind speed, relative humidity, incoming short wave radiation (all of which are shown in Figure 2.5), snow depth, long wave radiation and wind direction used in this study were recorded half-hourly from the three meteorological stations in Spring 2003 and from GB2 and P2 in Spring 2004 (Bewley, 2006; Pomeroy et al., 2006). Data from these periods are used in the thesis because, in addition to the meteorological stations, eddy covariance towers were set up at P2 and GB2 during the 2003 and 2004 snowmelt seasons. The sensible and latent heat fluxes, which were measured by finding covariances between vertical wind speed fluctuations and temperature or vapour density fluctuations respectively (Bewley et al., 2010), were used to evaluate model performance in Chapter
Figure 2.3: LiDAR based digital elevation map of the Granger basin overlaid with a canopy height map and marked with three transects (A, C and F) and with the location of two of the meteorological stations (GB2 and P2).

Figure 2.4: LiDAR based digital elevation map of the Granger basin overlaid by a vegetation fraction map.
7. There is little site-specific difference between the meteorological data except for the wind speed which is generally greater on the exposed plateau. Manual snow surveys, in which snow depths are measured every 5 m and snow density every 25 m with an ESC-30 snow tube (Figure 2.6-a) along three north to south transects of approximately 400 m, have been conducted every spring since 1998. Elevation along the three transects is shown Figure 2.7.

Snowfall for the 2003-2004 winter was assimilated by MacDonald et al. (2009) by extrapolating the snowfall measurements from the WIA and scaling it to GB with a multiplier determined from comparisons between manual snow surveys at GB and WIA. More details can be found in the above reference.

In order to evaluate the LiDAR data, manual vegetation distribution and canopy height surveys were conducted in August 2007 and in Spring 2008 (Figure 2.6-b). Figure 2.8 shows manual against LiDAR measurements along the A, C and F transects. As the measurements obtained with the two methods show reasonable agreement, the LiDAR data are used in the model used in Chapter 7 and Chapter 8 as input for the vegetation characteristics in the basins.

Figure 2.9 shows the frequency of winds in each direction in GB. The prevailing south to south-westerly winds causes a topographically-driven snow drift where snow blown from the plateau is deposited to occur at the top of the north-facing slope. Permafrost underlies most of this slope and the high elevations whereas seasonal frost dominates the south facing slope (Quinton et al., 2004).
Figure 2.5: Incoming shortwave radiation, air temperature, specific humidity and wind speed at P2 (red) and GB2 (black) from 15 April to 15 May 2004.
Figure 2.6: (a) Snow core taken with the ESC-30 snow tube to measure snow density. (b) Manual canopy height survey in Spring 2008.

Figure 2.7: Elevation along the A, C and F transects at GB. The transects start on the north-facing slope.
Figure 2.8: Manual (crosses) and LiDAR (triangles) canopy height measurements along three transects in the Granger basin
2.2.2 The alpine and the tall shrub sites

The meteorological station at the alpine tundra site (AS) is situated at 1560 m.a.s.l. (Figure 2.2-d). Vegetation cover at the site consists mainly of grass, forbs and lichen with isolated patches of dwarf birch and willow (0.01 - 0.3 m). Approximately one fifth of the site is unvegetated (Janowicz et al., 2004). Lewkowicz and Ednie (2004) suggested that the site was situated within the widespread discontinuous permafrost zone (0.5 - 0.9 permafrost probability).

The meteorological station at the tall shrub or “Buck brush” site (BB) is situated 1300 m.a.s.l. within a sporadic discontinuous permafrost zone (Lewkowicz and Ednie, 2004). The vegetation is typically 0.4 - 3 m tall and consists of willow, sparse white spruce, dwarf birch and grass (Figure 2.2-e).

Meteorological data used in this study at both sites are for a one year period starting 1 August 1998 and include net radiation, air temperature, wind speed, air pressure and specific humidity. In the absence of continuous snowfall measurements, snowfall was assimilated from automatic depth measurements provided by ultrasonic depth gauges (Campbell Scientific SR50). Soil temperatures taken at 5 cm depth at the alpine site and at 11 cm depth at BB are used in Chapter 3 to evaluate the snow schemes in JULES and in Chapter 8 to study the impact of shrub encroachment on soil temperatures in the WCRB. Data for this period were chosen to evaluate the model because they constitute a whole year for which continuous high quality soil temperature and meteorological measurements are available at the two sites.
The alpine site and the tall shrub “Buck brush” site (BB) were chosen because of their proximity and differences in altitude and vegetation cover. During the one year period covered by the data, air temperatures at both sites are similar from November to February, but AS is generally 2°C colder than BB for the rest of the year (Figure 2.12 a-b). Annual average wind speed at the alpine site is greater by 2.3 m s\(^{-1}\) (Figure 2.12 c). Increased wind ablation and reduced trapping by shrubs gives lower snow depths at the alpine site despite similar snowfall. Differences in summer soil temperatures are small (0.4°C from June to August with an average of 7°C at the alpine site and 6.6°C at the buck brush site) but the largest difference occurs in March when average soil temperatures are -9°C at the alpine site but -4°C at the buck brush site (Figure 2.12-d).
Figure 2.10: LiDAR based digital elevation and canopy height maps of the Alpine site.

Figure 2.11: LiDAR based digital elevation and canopy height maps of the Buck brush (BB) site.
2.3 The Torne and Kalix rivers basin, Northern Scandinavia

The Torne and Kalix rivers have their source in the Scandes mountain range which culminate at Mount Kebnekaise (2117 m), the highest point in Sweden. This river system is subject to much scientific investigations as it is the largest unregulated (i.e. no dam or constructions influence the water discharge) and unpolluted in Europe (Andersson et al., 2006; Nilsson et al., 1989; Elfendahl et al., 2006) (Figure 2.14). The rivers are covered by ice from November to April.

Figure 2.12: Scatterplots of (a) air temperatures when snow-covered (b) air temperatures when snow-free (c) wind speed and (d) soil temperatures at the “Buckbrush” and alpine sites. The red line in plots (a), (b) and (c) is the 1:1 line.
(Andersson et al., 2006), leading to ice jams and floods at the beginning of the snowmelt season when the streamflow rapidly increases.

The Torne river is 510 km long and has a catchment area of 40240 km². The river takes its source at approximately 68°5’ N in Norway, flows through to Sweden and constitutes a natural frontier between Sweden and Finland from 67°2’ N to its mouth. Its catchment area contains more than 1770 lakes and 410 river sub-catchments (Alanne et al., 2005). The Kalix river is 450 km long, covering 17950 km² and runs south of and almost parallel to the Torne river. The Torne and Kalix rivers are connected by a natural bifurcation leading to about 56% of the water from the northern parts of the Torne running to the sea through the Kalix (Elfvendahl et al., 2006)

There is an elevation gradient from the South-East of the basin, where the lowest point is at sea level and the rivers empty in the Gulf of Bothnia, to the North-West, in the Scandes mountain range (Figure 2.13). This gradient is also reflected in the annual precipitation which is dominated by orographic lifting and spans from 500 to 1600 mm per year. Snowfall forms on average 43% of the precipitation, ranging from 34% in the southeast to 56% in the northwest (Nijssen et al., 2003). Both elevation and latitude govern the vegetation cover. The South consists primarily of needle leaf boreal forests interspersed with large open areas of low relief mire and bog throughout the central portion of the basin (Bowling et al., 2003a). Further north, the landscape, then governed by elevation and latitude, transitions from shrub forests (predominantly birch) to an open shrub-tundra landscape (Figure 2.14).

The Torne and Kalix rivers basin is hereafter referred to as TKB.

2.3.1 Ovre Abiskojokk

The Ovre Abiskojokk catchment (hereafter referred to as Abisko) is a sub-basin within the Torne and Kalix rivers system (the northwesternmost basin in Figure 2.14). It is 566 km² with elevation ranging from approximately 350 m.a.s.l. to the highest point in TKB at 2117 m. It is situated at the head of the river Torne and water from the catchment outlets at lake Tornetrask, the largest within the TKB (Lyon et al., 2009). The catchment is situated in a zone of discontinuous permafrost with permafrost occurrence increasing with altitude and latitude although local permafrost can also be found on north facing slopes (Brown et al., 1988; Johansson et al., 2006). The alpine region is dominated by dwarf heath shrubs on shallow soils and by exposed bedrocks. The subalpine region is characterised by closed or open birch forests on deeper soils (Lyon et al., 2009).
Figure 2.13: Elevation (m) in the Torne and Kalix river basins. The gridboxes are 0.25° latitude x 0.25° longitude.

Mean annual temperature is approximately -2°C. The basin benefits from extensive long-term monitoring as it is host to the Abisko Scientific Research station (68°21’N, 18°49’E) first established in the late 19th century. It is of particular interest for this study as it is situated in the shrub-tundra ecozone, where the dominant shrub species is *Betula pubescens.* Meteorological data, averaged over 1989-1998, for the gridbox in the Abisko basin where the Abisko Scientific Research station is situated, are shown in Figure 2.15.
Figure 2.14: Land cover of the Torne and Kalix river basins. Taken from Bowling et al. (2003b).
Figure 2.15: 10 years hourly average (1989-1998) meteorological data for one gridbox (0.25° latitude x 0.25° longitude) in the Abisko catchment. From top to bottom: Incoming shortwave radiation, air temperature, specific humidity, wind speed and cumulative snowfall (solid line) and rainfall (dash line).
2.3.2 Atmospheric forcing, evaluation data and ancillary files

Atmospheric forcing and hydrological data for the TKB, provided by Dr Laura Bowling, Purdue University, and originally compiled for the PILPS 2(e) experiments (Bowling et al., 2003a), were used in this thesis because they provide quality controlled gridded data at high latitude. To the author’s knowledge, it is a unique dataset because it encompasses meteorological and hydrological data over a 20-year period in a shrub-tundra basin, where harsh winter conditions make it difficult to maintain continuous reliable data (Bowling et al., 2003a). The Swedish Meteorological and Hydrological Institute provided daily discharge and meteorological station data from 01/01/1979 to 31/12/1998 for the inter-comparison project, as well as a 1° dataset (Mueller, 2003) from which other key meteorological parameters were interpolated (e.g. air pressure, cloud cover) for use in PILPS 2(e).

The TKB area was distributed into 218 1/4° gridboxes which at 67°N corresponds to cells of approximately 28 km in longitude and 11 km in latitude (Nijssen et al., 2003). The land surface model JULES (Joint UK Land Environment Simulator) used in Chapters 3, 4 and 6 was driven by precipitation (rainfall and snowfall), air temperature, surface pressure, specific humidity, wind speed and incoming longwave and shortwave radiation. Some land-surface characteristics were also provided for parameterization:

- Basin boundaries for the Torne and Kalix river basins as well as the Abisko sub-basin.
- Soil textural and soil bulk densities for the top metre of soil.
- Vegetation type, height and albedo.
- Leaf Area Index (LAI)

Model performance was evaluated against snow cover and daily discharge. The Northern Hemisphere EASE-GRID Weekly Snow Cover and Sea Ice Extent 25 x 25 km product (Armstrong and Brodzik, 2002) was used, as in PILPS 2(e) (Nijssen et al., 2003), to evaluate the model against the first remotely sensed snow-free day. This product is derived by Professor David Robinson, Rutgers University, from the NOAA-NESDIS Weekly Northern Hemisphere Snow Charts which cover 125 x 205 km (Rutgers University Global Snow Lab, 1999). Streamflow was evaluated against measurements obtained from 3 gauged stations: two situated at the mouths of the Torne and the Kalix rivers and one downstream from the Abisko basin (Bowling et al., 2003a).
2.4 Summary

In this chapter, a description of the Wolf Creek Research Basin, Yukon Territory, Canada, and the Torne and Kalix river basin, Fennoscandia, were presented. Further details on the sub-catchments of the two basins situated within the shrub-tundra ecozone and on the data used to inform the initial conditions of the models and to perform model evaluation were introduced.
In the next chapter, the first model used in this study to investigate the effect of shrub-tundra on the hydrology in the TKB, JULES, is introduced.
Chapter 3

Description of the physics in the Joint UK Land Environment Simulator (JULES)
3.1 Introduction

JULES (Joint UK Land Environment Simulator) is a UK community land surface model originally based on the code of the Met Office Surface Exchange System (MOSES), which was developed for use in meteorological and climate models, notably the Met Office Unified Model (UM). When JULES was first launched, its internal process representation was identical to that of MOSES but recent work, some of which is discussed below, has added new functionality to JULES. Detailed descriptions of the physics relevant to this study are in the following sections but descriptions of the UM and MOSES can be found in Gregory et al. (1994), Cox et al. (1999) and Essery et al. (2001, 2003a).

3.2 Model Structure

A simple schematic representation of the model is presented Figure 3.1. The model solves the water and energy balances at the Earth’s surface. Landcover subgrid heterogeneity is represented by the energy balance being solved separately over different surface types (or tiles) which interact independently with the atmosphere at a reference height. On the other hand, each gridbox comprises a single soil column, meaning that all the surface types share the same soil characteristics and that a single soil water balance is calculated for the gridbox.

JULES 2.1 was developed to answer the needs of a heterogeneous community of scientists investigating a wide range of land surface processes. Since its launch in 2006, JULES has been used to investigate a range of meteorological and biogeochemical processes such as the spatial pattern of thermal microclimate in a UK upland (Bennie et al., 2010), the ecosystem productivity in Africa (Weber et al., 2009) and in boreal (Alton et al., 2007; Prieto-Blanco et al., 2009), temperate and Amazonian forests (Alton et al., 2007), the European carbon balance (Harrison et al., 2008) and the surface energetics of patchy Arctic snowcover (Wiltshire, 2006). Although there exist other models that are constantly being developed to improve the representation of land surface processes (e.g. CLASS, Verseghy, 1991; ISBA, Douville et al., 1995; ECHAM, Roesch et al., 2001; CLM, Oleson et al., 2004) few can address as wide a range of issues within global change research. In addition, the structure of the model was developed with the intention of providing a number of flexible features. Of most relevance to the present study are the following, some of which were already available in MOSES:

* the explicit representation of subgrid heterogeneity within a gridbox through
Figure 3.1: Simple representation of the energy and water balances in JULES where $SW$ and $LW$ is net short and long wave radiation respectively, $LE$ is latent heat flux and $H$ is sensible heat flux. Details of the surface when snow is present are found in Figure 3.3.

- a flexible tile scheme. Each tile is characterised by different surface types or sub-areas which are defined by biophysical parameters. The UM implementation of MOSES comprises 9 tiles: broadleaf trees, needleleaf trees, $C_3$ (temperate) grass, $C_4$ (tropical) grass, shrubs, urban, bare soil, inland water and ice.

- user defined number and depth of soil layers. The default soil configuration implemented in MOSES and used in JULES for this study comprises 4 soil layers of thickness 0.1 m, 0.25 m, 0.75 m and 2.00 m. This configuration was chosen to capture diurnal and seasonal variation in the surface evaporation and runoff fluxes and to enable gradients in soil water tension between layers to be modelled (Gregory et al., 1994).

- the driving of the model over a grid or at a point. There are no restrictions on grid box size although the absence of horizontal interactions within and
between cells would lead to poor physical representation of surface processes at too small or too large a scale.

- the driving of the model being provided by an atmospheric model (online) or by measured meteorological data (offline). JULES requires rainfall, snowfall, air pressure, specific humidity, air temperature, wind speed, incoming shortwave and incoming longwave radiation for model running. All simulations in this study will be performed offline.

- the flexible multi-layer snow scheme. The model can be run with a single composite snow layer or a separate user-defined number of snow layers. More details about the snow scheme are provided in Section 3.7.

3.3 The surface energy balance

Following the convention that radiative fluxes are positive toward the surface, and turbulent and ground heat fluxes are positive away from the surface, the energy balance for each tile is

\[ R - H - LE - G_0 - L_f S = 0, \]  

(3.1)

where \( H, LE \) and \( G_0 \) are the sensible, latent and ground heat flux respectively, \( L_f \) is the latent heat of fusion and \( S \) is the snowmelt rate. Net radiation, \( R \), for surface emissivity assumed to equal 1, is obtained as

\[ R = (1 - \alpha_{tile})SW + LW - \sigma T^4, \]  

(3.2)

where \( \alpha_{tile} \) is the tile albedo, \( SW \) and \( LW \) are incoming shortwave (0.15 to 3 \( \mu \)m) and longwave (3 to 100 \( \mu \)m) radiation respectively, \( T \) is the surface skin temperature and \( \sigma \) is the Stefan-Boltzmann constant. Linearising about surface soil layer temperature, \( T_{ss} \), gives

\[ R = R_{ss} + 4\sigma T_{ss}^3(T_{ss} - T) \]  

(3.3)

for

\[ R_{ss} = (1 - \alpha_{tile})SW + LW - \sigma T_{ss}^4. \]  

(3.4)

\( H \) is calculated as

\[ H = C_p \frac{\rho}{T_a} (T_a - T). \]  

(3.5)
where \( \rho \) and \( C_p \) are the density and the heat capacity of air respectively, \( r_a \) is the aerodynamic resistance and \( T_a \) is the air temperature at reference height. The moisture flux, \( E \), is calculated as

\[
E = \psi \frac{\rho}{r_a} \left[ Q_{sat}(T_*, P_*) - Q_a \right],
\]

(3.6)

where \( \psi \) is a soil moisture availability factor, \( Q_{sat}(T_*, P_*) \) is the saturation humidity at temperature \( T_* \) and surface pressure \( P_* \) and \( Q_a \) is the specific humidity at the reference height. \( LE \), in Equation 3.1, is the product of the moisture flux and the latent heat of condensation, \( L_c \), for an unfrozen surface or the latent heat of sublimation, \( L_s \), for a frozen surface. \( G_0 \) is parameterized as

\[
G_0 = \frac{2 \lambda}{\Delta z_1} (T_* - T_{ss}),
\]

(3.7)

where \( \lambda \) and \( \Delta z_1 \) are the thermal conductivity and thickness of the top soil layer respectively.

Substituting Equations 3.3 and 3.7 in Equation 3.1 gives the surface temperature as

\[
T_* = T_{ss} + \frac{1}{A_*} (R_{ss} - H - LE - Lf S_M),
\]

(3.8)

where

\[
A_* = \frac{2 \lambda}{\Delta z_1} + 4 \sigma T_{ss}^3.
\]

(3.9)

Linearizing \( Q_{sat} \) about \( T_a \) to get the humidity deficit, \( \Delta Q_a \), and using Equation 3.8 to eliminate \( T_* \) from Equations 3.5 and 3.6 by using an extended version of the Penman-Monteith approach gives

\[
E = \psi \frac{\rho}{r_a} \left[ \frac{D(\tilde{R} - Lf S_M) + (C_p \rho/r_a + A_*) \Delta Q_a}{(C_p + LD \psi) \rho/r_a + A_*} \right],
\]

(3.10)

and

\[
H = C_p \frac{\rho}{r_a} \left[ \frac{\tilde{R} - Lf S_M - L \psi (\rho/r_a) \Delta Q_a}{(C_p + LD \psi) \rho/r_a + A_*} \right],
\]

(3.11)

where \( D \) is the slope of saturation vapour pressure, \( L \) is \( L_c \) or \( L_s \) and \( \tilde{R} \) is

\[
\tilde{R} = R_{ss} - A_*(T_a - T_{ss}).
\]

(3.12)

When snow is present, \( S_M \) in Equation 3.1, 3.10 and 3.11 is set to 0 to give a first estimate of the surface temperature. If the diagnosed \( T_* \) exceeds the melting point, \( T_m \), the residual in the energy balance is used to melt snow.
3.4 The albedo parameterization

To represent the albedo of a heterogeneous surface, the tile albedo is calculated by weighting the albedo of the surface type with that of the snow such that

\[ \alpha_{\text{tile}} = F_s \alpha_s + (1 - F_s) \alpha_{\text{st}} \]  \hspace{1cm} (3.13)

where \( \alpha_{\text{tile}}, \alpha_s \) and \( \alpha_{\text{st}} \) are the tile, snow and snow-free surface albedos and \( F_s \) is the snow cover fraction. The calculation of the snow cover fraction differs whether the model is run at a point or over a gridbox (Figure 3.2). This is to account for the fact that at a point snow is either present or not, whereas the area covered by a gridbox becomes snow covered more progressively. For single point simulations, \( F_s \) is calculated as

\[ F_s = 1 - e^{-50S_d} \]  \hspace{1cm} (3.14)

where \( S_d \) is the snow depth, and for distributed runs

\[ F_s = \frac{S_d}{S_d + 10z_0} \]  \hspace{1cm} (3.15)

where \( z_0 \) is the momentum roughness length (Essery et al., 2003a).

Many LSMs include a snow cover parameterization derived from their predicted SWE or snow depth to simulate sub-grid heterogeneity, although formulations differ between models (e.g. see Verseghy, 1991, for CLASS; Douville et al., 1995, for ISBA; Roesch et al., 2001, for ECHAM). In JULES, the effect of vegetation on snow cover is implicit in the use of the roughness length in Equation 3.15 such that, for equal snow depth, a tall vegetation tile is diagnosed with a smaller \( F_s \) than a short vegetation tile and, by extension, has a lower albedo.

3.5 Soil hydrology and thermodynamics

JULES uses a heat conduction approach to calculate soil temperatures in each layer. The rate of change of temperature in layer \( n \) of thickness \( \Delta z_n \) is

\[ C_A \Delta z_n \frac{dT_n}{dt} = G_{n-1} - G_n - J_n \Delta z_n \]  \hspace{1cm} (3.16)

where \( G \) and \( J \) are the diffusive and advective fluxes in and out of the layer and \( C_A \) is the volumetric heat capacity of the layer, which is dependent on the amount of dry soil, liquid water and ice and their associated phase changes (Cox et al., 1999).
The fluxes are calculated from gradients between the middle of the layers. The diffusive and advective fluxes are given by:

$$ G = \lambda \frac{\partial T}{\partial z} $$  \hspace{1cm} (3.17)

and

$$ J = c_w W \frac{\partial T}{\partial z}, $$  \hspace{1cm} (3.18)

where $\lambda$ is the thermal conductivity, $c_w$ is the specific heat capacity of water and $W$ is the vertical moisture flux.

The total soil moisture content of a layer, $S_n$, is calculated as

$$ S_n = \rho_w \Delta z_n \theta_{sat} \Theta_n $$  \hspace{1cm} (3.19)

where $\rho_w$ is the water density, $\theta_{sat}$ is the volumetric saturated soil moisture content and $\Theta_n$ is the total soil moisture content as a fraction of $\theta_{sat}$.

The change in storage, $\delta S_n$, in each layer is updated every timestep such that

$$ \frac{\delta S_n}{\delta t} = W_{n-1} - W_n - E_n $$  \hspace{1cm} (3.20)
where $\delta t$ is the time increment, $W$ is the water flux and $E_n$ is the evapotranspiration extracted directly from the layer by plant roots. 

$W$ between layers is calculated using a finite difference approximation to the Richards’ equation which combines Darcy’s law for vertical flow of water in an unsaturated soil with the conservation of mass such that

$$W = K(\Theta) \frac{d[z + \Psi(\Theta)]}{dz}$$

(3.21)

where $K$ is the hydraulic conductivity and $\Psi$ is the soil water suction. The relationship between matric potential, moisture content and hydraulic conductivity is calculated using the Clapp and Hornberger (1978) dependencies where

$$\Psi = \Psi_{sat} \Theta_u - b$$

(3.22)

$$K = K_{sat} \Theta_u^{2b+3}$$

(3.23)

where $\Psi_{sat}$ and $K_{sat}$ are the saturated soil water suction and hydraulic conductivity at saturation respectively, $b$ is an empirical soil-dependent constant and $\Theta_u$ is the unfrozen water content as a fraction of saturation. The relationship of soil moisture characteristics to the physical properties of soils is calculated using empirical dependencies of the soil properties based on Cosby et al. (1984).

### 3.6 Runoff generation

Rainfall, snowfall, snow melt, dew, frost and canopy throughfall constitute the water input at the surface. Large scale and convective precipitation is assumed to be exponentially distributed in space and to fall on a fractional area within the gridbox (Gregory et al., 1994). The default values are 1 for large-scale and 0.3 for convective precipitation, although they can be changed to better suit gridbox scale.

Surface and subsurface runoff are commonly thought of as fast and slow hydrological responses respectively. Water infiltrates a soil where it is either stored, diagnosed as surface runoff or passed to the layer below according to Equation 3.20.

Surface runoff in JULES can be generated as infiltration excess overland (or Hortonian; Horton, 1945) flow which occurs when water input rate exceeds surface infiltration capacity. The surface infiltration rate is the product of the saturated hydraulic conductivity and a tile specific enhancement factor that accounts for
the effects of root systems on the infiltration of surface water into the soil. Surface runoff can also occur to avoid supersaturation of a soil layer. If soil moisture in a layer exceeds the saturated soil moisture content, water in excess of saturation is forced to the layer directly above the saturated layer. The excess moisture is translated into surface runoff if the saturated layer is the top soil layer. Subsurface runoff occurs at the bottom boundary. Equation 3.21 is then reinterpreted with $\partial \Psi / \partial z = 0$, corresponding to free or gravitational drainage, whereupon the outflow rate from the bottom layer equals the hydraulic conductivity (Cox et al., 1999).

### 3.7 The snow schemes

A choice of two snow schemes is offered in JULES 2.1: the Composite soil / snow layer scheme (CL) and the Multiple snow layer scheme (ML). The simpler CL predates developments in LSMs motivated by intercomparison projects like PILPS and SNOWMIP and was implemented in the UM (Smith, 1993) to inform the atmospheric model of the conditions at the surface. Since then, the land surface community has recognised the importance of model structure to the representation of snowmelt processes (e.g. Slater et al., 2001; Etchevers et al., 2004) and more complex schemes, such as ML, that allow better representations of the structure of the snowpack and of snow physics have been introduced.

#### 3.7.1 The composite soil / snow layer

When snow is present, the top soil layer becomes a soil / snow composite that thermally functions as a single layer. The temperature of this layer is taken at a fixed depth below the surface ($\Delta z_1 / 2$) whether snow is present or not, hence representing the soil or the snow temperature depending on snow depth ($S_d$). The insulating effect of the snowpack is accounted for by reinterpreting the thermal conductivity of the layer, $\lambda_{CL}$, as that of snow when $S_d > \Delta z_1 / 2$ (as in Figure 3.3a) or as a combination of snow and soil when $S_d < \Delta z_1 / 2$ (as in Figure 3.3b) such that

$$\lambda_{CL} = \lambda \left[ \frac{\Delta z_1}{\Delta z_1 + 2S_d(\lambda/\lambda_s - 1)} \right]$$ (3.24)

where $\lambda_s$ is thermal conductivity of snow. The heat flux at the bottom of the surface layer, given by Equation 3.17, is reduced by a multiplicative factor to
account for the increased distance between the soil layers and the surface given by

\[
\zeta = \frac{\Delta z_1 + \Delta z_2}{\Delta z_1 + \Delta z_2 + 2S_d}
\]  

(3.25)

for \(S_d \leq 0.5\Delta z_1\) and

\[
\zeta = \frac{\Delta z_1 + \Delta z_2}{2\Delta z_1 + \Delta z_2 + \lambda / \lambda_s (2S_d - \Delta z_1)}
\]  

(3.26)

for \(S_d > 0.5\Delta z_1\). The thermal conductivity and density of snow are fixed user-defined parameters whose default values are 0.265 W m\(^{-1}\) K\(^{-1}\) and 250 kg m\(^{-3}\) respectively. Retention and freezing of liquid water in snow are neglected.

Figure 3.3: Schematic representation of the 2 snow schemes in JULES 2.1. Both (a) and (b) show the soil / snow composite layer, (a) when the diagnosed snow depth > 0.5\(\Delta z_1\), (b) when the diagnosed snow depth < 0.5\(\Delta z_1\) and (c) shows the flexible multi-layer snowpack.
3.7.2 The multi-layer snow scheme

A flexible multiple snow layer structure is distinguished from the soil model, separating the thermal regime of the snow from that of the soil (Figure 3.3-c). ML was introduced in JULES 2.1 by Richard Essery, University of Edinburgh, Douglas Clark, Centre for Ecology and Hydrology, and Martin Best, Met Office. Snow layer depths and properties are updated at each timestep, allowing a density profile to develop with less-dense freshly fallen snow in the surface layer and mechanical compaction leading to higher snow densities in deeper layers. Upon growth of the snowpack, the first snow layer increases in thickness until it reaches a prescribed depth, at which point the layer splits in two. The thickness of the top snow layer then stays fixed and subsequent increases in snow depth are accommodated by increasing the depth of the new lower layer. This process is repeated for each subsequent layer as snow accumulates until a prescribed maximum number of layers is reached. Separate thermal conductivities and densities are calculated for each layer and updated at each timestep. Upon snowmelt, the shallowing of the snowpack is accommodated by decreasing the depth of the deepest snow layer first. The model reverts to the snow scheme described in Section 3.7.1 when $Sd$ is less than the prescribed surface snow layer thickness to avoid numerical instability.

Following extensive measurements in Hokkaido, Kojima (1967) proposed a model for the compaction of snow such that

$$\frac{\delta \rho_s}{\delta t} = \rho_s \frac{\sigma}{\eta},$$

(3.27)

where $\rho_s$ is snow density, $\delta t$ is a time increment, $\sigma$ is stress and $\eta$ is compactive viscosity. Stress on layer $n$ is defined as the load exerted by the snowpack on its base such that

$$\sigma = g S_{\text{mass}},$$

(3.28)

where $g$ is the gravitational acceleration and $S_{\text{mass}}$ is overlying mass given by

$$S_{\text{mass}} = 0.5(I_n + W_n) + \sum_{i=1}^{n-1} (I_i + W_i),$$

(3.29)

for layer ice and water mass contents $I$ and $W$. $\eta$ depends on both the density and the temperature. Kojima (1967) gives the viscosity in the temperature range 268 - 273 K, $\eta_0$, as
\[ \eta = \eta_0 \exp \frac{\rho_s}{\rho_0} \]  
where \( \rho_0 \) is 50 kg m\(^{-3}\). The dependency of \( \eta \) on the temperature of the snow layer, \( T_s \), is given as

\[ \frac{\Delta (\ln \eta)}{\Delta (1/T_s)} = k_s. \]  
(3.31)

Kojima (1967) established \( \eta_0 \) to be 10 MPa s for \( \rho_0 \) and \( k_s \) to be 4000 K. Substituting Equation 3.30 into Equation 3.31 and reorganising to get \( \eta \) gives

\[ \eta = \eta_0 \exp \left( \frac{\rho_s}{\rho_0} + \frac{k_s}{T_s} - \frac{k_s}{T_m} \right). \]  
(3.32)

Substituting Equations 3.28 and 3.32 in Equation 3.27 and accounting for the addition of fresh snow, with density \( \rho_{\text{new}} = 100 \text{ kg m}^{-3} \), gives

\[ \delta \rho_s = 10^{-7} \rho_s g S_{\text{mass}} \delta t \exp \left( 14.643 - \frac{4000}{T_s} - 0.02 \rho_s \right) + \frac{S_f(\rho_{\text{new}} - \rho_s)}{S_{\text{mass}}}, \]  
(3.33)

where \( S_f \) is snowfall added to the layer. The thermal conductivity of the layer is taken from Yen (1981) as

\[ \lambda_s = \lambda_i \left( \frac{\rho_s}{\rho_{\text{water}}} \right)^{1.88}, \]  
(3.34)

where \( \lambda_i = 2.22 \text{ W m}^{-1} \text{ K}^{-1} \) is the thermal conductivity of ice and \( \rho_{\text{water}} = 1000 \text{ kg m}^{-3} \) is the density of water. Surface skin temperature is diagnosed from the surface energy balance and is not allowed to exceed 0°C. If a snow layer is diagnosed as being above the melting point, the layer ice mass, \( I_n \), is reduced by

\[ \Delta I_n = \frac{C_n}{L_f} (T_s - T_m), \]  
(3.35)

where \( C_n \) is the areal heat capacity of the layer \( n \) calculated as

\[ C_n = I_n C_{\text{ice}} + W_n C_{\text{water}} \]  
(3.36)

for \( C_{\text{ice}} \) being the specific heat capacity of ice, \( C_{\text{water}} \) being the specific heat capacity of water and \( W_n \) being the water content in the snow layer. The liquid water mass of the layer is increased by the same amount as \( \Delta I_n \) and \( T_s \) is reset to 0°C. If the user-defined maximum \( W_n \) is reached, excess water is moved to the layer below. When it reaches the bottom of the snowpack, \( W_n \) is passed as snowmelt to the soil column where it is partitioned between infiltration and
surface runoff. Refreezing of $W_n$ and its resulting latent heat release occurs when water flows to a layer where $T_s < T_m$.

### 3.7.3 The canopy model

JULES includes an optional canopy model, based on models by Hedstrom and Pomeroy (1998) and Pomeroy et al. (1998), whereby snow interception at the branch scale is applied to the whole canopy. The model was introduced in MOSES by Essery et al. (2003b).

The intercepted snow mass, $I$, decreases over time due to unloading according to

$$ I = I_i e^{-Ut} $$

where $U$ is a snow unloading coefficient, $t$ is time since previous snowfall and the interception following snowfall, $I_i$, is

$$ I_i = (I_{\text{max}} - I_0)(1 - e^{S_f/I_{\text{max}}}) $$

where $I_0$ is the initial load and $I_{\text{max}}$ is the maximum snow load that the canopy can sustain. Following Schmidt and Gluns (1991),

$$ I_{\text{max}} = \overline{S}(0.27 + 46/\rho_s)\text{LAI} $$

where $\overline{S}$ is a species-specific parameter. From measurements taken on conifer trees and for typical values of $\rho_s$ and $\overline{S}$, Schmidt and Gluns (1991) give $I_{\text{max}} = 4.4\text{LAI}$ where LAI is the leaf area index. Hedstrom and Pomeroy (1998) determined $e^{-Ut}$, from measurements of $I$ and model simulations of $I_i$, to have a mean value of 0.678.

In addition to unloading, intercepted snow may be removed from the canopy by sublimation or melt. When the canopy model is used, sublimation occurs from the intercepted snow but the formulation of the moisture flux neglects sublimation from snow on the ground such that Equation 3.6 is reinterpreted as

$$ E = \frac{\rho}{r_a + r_i}[Q_{\text{sat}}(T_c, p_s) - Q_a] $$

where $r_i$ is the resistance for transport of moisture from the intercepted snow to the canopy air space. Pomeroy et al. (1998) and Essery et al. (2003b) adapted the sublimation of ice spheres model from Thorpe and Mason (1966) for application on canopy snow to obtain a resistance
\[ r_i = \frac{2\rho_i r^2}{3C_e I D_w Sh}, \]  
\[ (3.41) \]

where \( \rho_i \) is the density of ice, \( r \) is a nominal grain size, \( C_e \) is a snow grain exposure coefficient, \( D_w \) is the diffusivity of water vapour in air and \( Sh \) is the Sherwood number, a ratio of convective to diffusive mass transport.

The canopy is treated as opaque so there is no penetration of shortwave radiation through the canopy to the ground. Allowing for longwave radiation between the ground and the canopy gives a modified version of Equation 3.2 as

\[ R_c = (1 - \alpha_{tile})SW_i + LW_i - \sigma T_{ss}^4 - 2\sigma T_c^4 \]
\[ (3.42) \]

where \( R_c \) is the net radiation absorbed by the canopy and \( T_c \) is the canopy temperature. \( H \) beneath the canopy is

\[ H_c = C_p \frac{\rho}{r_{ah}} (T_{ss} - T_c) \]
\[ (3.43) \]

where \( r_{ah} \) is an in-canopy resistance. The energy balance of the snow-covered canopy is then given by

\[ C_c \frac{dT_c}{dt} = R_c - H - H_c - LE - L_f S_M \]
\[ (3.44) \]

where \( C_c \) is an areal canopy heat capacity, calculated assuming specific heat capacities (in kJ K\(^{-1}\) kg\(^{-1}\)) of 570 for leaves, 110 for wood and 2.1 for snow (Essery et al., 2003b).

### 3.7.4 Evaluation of the two snow schemes

The performance of the two snow schemes is evaluated against measured snow depth, soil temperatures and snow densities at the alpine and Buckbrush sites, for which descriptions can be found in Chapter 2. The snow densities used to validate the model are monthly climatological densities calculated from 10 years of snow surveys provided by the Government of Yukon. Model evaluation against snow depths and soil temperatures taken at 5 cm depth at the alpine site and at 11 cm depth at BB are presented in Figure 3.4. Modelled soil temperatures correspond to the soil temperature in the middle of the layer. While the depth at which soil temperatures were measured at the alpine site coincides with the middle of \( \Delta z_1 \), temperature depth at BB coincides with the top of \( \Delta z_2 \). As a consequence, modelled soil temperature at BB was depth weighted between \( \Delta z_1 \) and \( \Delta z_2 \).
The structure of the snowpack with CL leads to underestimation of winter soil temperatures not only because the surface soil temperature lies within $\Delta z_1$ but also because the thermal conductivity is overestimated by the fixed snow density. On the other hand, the insulation provided by the separate snowpack with ML considerably improves soil temperature simulations. A bias towards cold temperatures still occurs in October at both sites, reflecting the fact that ML = CL for shallow snow.

The implementation of varying snow density also improves the snow depth simulations. Allowing for less dense fresh snow and for compaction allows the model to reproduce the snow depth accurately during snow accumulation. This mechanism is illustrated in Figure 3.5 where 10-year average monthly densities calculated from snow surveys are plotted against CL and ML densities. While ML provides a good fit to climatological measurements, the fixed CL snow density is always greater than measurements, leading to a shallower snowpack, as can be seen in Figure 3.4 c-d.
Figure 3.5: Measured climatological (squares) snow density at (a) BB and (b) AS against fixed (CL, black line) and modelled (ML, green line) values with JULES. Error bars show observed monthly standard deviation.

3.8 Summary

Land surface models are powerful tools that allow the investigation of processes at the land surface. Although they were historically implemented to provide boundary conditions for atmospheric models, developments over the past 20 years have lead to more complex, physically accurate LSMS. In JULES, the most recent development is the implementation of a separate multiple snow layer scheme which allows a better representation of the physics and properties of the snowpack. By separating the thermal regime of the snow from that of the soil, the new snow scheme now allows JULES to be used in the investigation of subsurface processes in cold regions. In the next chapter, the effect of snowmelt on surface and subsurface runoff in a high latitude basin will be investigated and alternative formulations of soil processes will be proposed in order to improve the performance of the model at high latitudes.
Chapter 4
Improving the hydrological processes at high latitudes in JULES
4.1 Introduction

Runoff is projected to increase by 10 to 40% at high latitude by the middle of the century (Kundzewicz et al., 2007). On land, such changes are expected to shift seasonal runoff and affect water supply, water quality and increase the risk of floods. In the oceans, the most extreme scenarios suggest that the thermohaline circulation could shut down which would lead to dramatic changes in the global climate (Broecker, 1997). As a consequence, it is essential that land surface models are able to represent the hydrological processes at high latitudes.

Arctic hydrology is dominated by the annual cycle of snow: precipitation is stored in the landscape for up to nine months and is released over just a few days to weeks during the spring snowmelt period. Permafrost distribution, seasonal active layers and organic soils exert dynamic controls on the distribution, the movement and the storage of water (Hinzman et al., 2005). During spring snowmelt, some areas can generate surface flow when snow melts over an impermeable frozen layer while others can promote sub-surface hillslope hydrological responses if they are underlaid by highly conductive organic soils. The challenge for LSMs is therefore to include such critical hydrological processes within large-scale models while being computationally efficient.

Here, two schemes aimed at simulating the horizontal sub-grid heterogeneity of the soil processes in JULES are proposed. One other scheme which modifies the treatment of soil moisture in excess of saturation is also presented. In this chapter, two terms are used to described model output. Firstly, the term *runoff* describes the water that is produced in each gridbox as surface or subsurface discharge. Secondly, the term *streamflow* is used to describe surface or subsurface discharge which has been routed through a river network, thus allowing for a delay between water production in headwaters and flow at the outlet point. More details can be found in Section 4.3. All schemes are evaluated against streamflow at the outlet point of the Abisko catchment and of the Torne and Kalix rivers.

The first 10 years of the data (1979-1988) were used as a spin-up period, *i.e.* used as a phase during which the model is expected to reach a quasi-equilibrium state, and were not included in the results. All figures in this section are averages of the last 10 years of the data (1989-1998) unless stated otherwise. This averaging was deemed to be adequate for the purpose of this study because measured streamflow in the basins show the same broad patterns over the years: low late autumn and winter flows and high snowmelt-induced springtime discharge followed by rainfall-induced summer and early autumn flow.
4.2 Evaluation of the snow cover simulation

Before looking at runoff and streamflow simulations at high latitudes, it is essential to assess whether the model is able to capture correctly the timing of snowmelt, which dominates the hydrology of nival regimes. The Northern Hemisphere EASE-GRID Weekly Snow Cover and Sea Ice Extent products (Armstrong and Brodzik, 2002) are used to compare the observed (Figure 4.1-a) and the modelled (Figure 4.1-b) first snow-free day (SFD) of the year.

4.2.1 Issues with the remote sensing product

Simulations for the PILPS2(e) experiments revealed that the two versions of MOSES being assessed melted the snow too early, sometimes up to four to six weeks before observations (Nijssen et al., 2003). However, model output and observations were compared over different scales as only 6 satellite-based images cover the 218 cells of the TKB (Figure 4.2). Indeed, although the resolution of the final product is 25 x 25 km, the original product range is 125 x 205 km (see Section 2.3.2). Rather than provide more information on snow cover than the original snow charts, the products used in PILPS2(e) and in this study are simply regridded on a 25 x 25 km grid (Nijssen et al., 2003).

Another issue with the satellite-based data is the definition of a snow-free pixel. In PILPS2(e), the first snow-free day in each year was taken to be the first day that the reported average snow water equivalent in a model grid cell was less than 10 mm. A non-zero threshold was selected because the satellite product classifies a pixel as snow-free when more than 50% is snow free (RUGSL, 1999). Thus, two intrinsically different snow properties - satellite-based aerial snow extent and modelled snow water equivalent - were compared to evaluate the model.

Finally, Nijssen et al. (2003), who only had data available from 1989 to 1995 in PILPS 2(e), identified disagreements between ground-based measurements and the satellite product. The latter showed that snow disappeared in May and reappeared in June in the northwest of the basin, but the former did not. Nijssen et al. (2003) concluded that the difference could be an artefact of the classification procedure and, therefore, did not use the product in 1994 and 1995, when the difference was especially large. Although the remote sensing data are available until the end of the runs (1998) for this study, they were not used because the potential errors in the northwest of the basin were not quantified. As a consequence the model is evaluated against the satellite images from 1989-1993.

Notwithstanding these issues, the satellite data are useful to capture the broad
snow melt patterns across the TKB and are used, if only with caution, to evaluate the modelled first snow-free day.

Figure 4.3 shows the climatological SFD difference between the 208 model grid points and the 6 satellite images. Positive values denote that the model SFD occurs earlier than the observed SFD. The difference is considered in terms of weeks in order to use the same unit as the remote sensing product where “0” denotes that the modelled SFD falls within the same week as the observed SFD.

4.2.2 Results

54% of the modelled SFD falls in the same week or within 1 week of the observed SFD, including 40% in the southern half of the catchment. There is a clear distinction between the North where the model has a tendency to diagnose the SFD up to 5 weeks earlier than the remote sensing product and the South where the difference is mostly no more than 1 week late. Melt is systematically diagnosed early in the northwest where the difference with satellite imagery ranges from 1 to 5 weeks.

However, the largest model to observations SFD differences may reflect the difference in resolution between the model cells and the satellite images rather than model errors. For example, Figure 4.1-b shows that the model is able to capture the gradual advance of the SFD from south to north, persisting for the longest time in the northwest. While the observed SFD also captures this pattern (Figure 4.1-a), the gradient is less defined than in the model because the satellite images cover larger areas. Disagreements in SFD also occur in the mountainous region in the northwest of the basin where the model seems to capture some of the altitudinal range but where the larger scale of the remote sensing product cannot.

Differences in elevation, vegetation or meteorological conditions do not appear to explain why the model melts at 11 gridboxes near the mouth of the river four to five weeks later than their surrounding cells and five weeks later than the weekly snow cover product. Although 7 out of these 11 cells fall within a different elevation band to their surrounding (< 100 m against > 100 up to 200 m) it is unlikely to cause such differences and would not explain the SFD difference for the other 4 gridboxes. Further investigation into the timing of snowmelt in this particular area may identify the source of the difference. It may reflect a model error or an actual process that the model is able to reproduce but that the remote sensing products, because of their resolution, fail to recognise. As the main focus of this study is to investigate hydrological processes in the shrub-tundra ecozone
situated at the north of the basin, no further investigation into this difference in snowmelt occurring in the woodland in the south of the basin is conducted herein.

Point measurements of snow water equivalent or of snow depth in the basin or finer resolution satellite-based products are needed to provide more confidence in the model’s ability to simulate snow cover over the TKB. However, the comparison between the model and the available satellite-based products suggest that the model is able to reproduce the broad snow cover patterns in the basin.
Figure 4.1: Average (a) observed and (b) modelled snow free date for the 1989 to 1993 period in the Torne and Kalix river basin.
Figure 4.2: Mask of the 6 remote sensing images available for the area covering the Torne and Kalix catchment.

Figure 4.3: SFD difference in weeks (observations - model) for the 1989-1993 period.
4.3 The TRIP river routing scheme

In the absence of a routing scheme, modelled runoff from the contributing catchment area of each grid box can be aggregated then scaled with the upstream basin area for comparison with observed basin discharge. While this method is appropriate for

1. small basins where the runoff production is quickly translated into streamflow
2. calculating average runoff production over time intervals that are longer than the time taken for the runoff to be translated into streamflow

it is not applicable for shorter time intervals and larger basins, when the lag between runoff production and streamflow is more important.

The simple river routing scheme Total Runoff Integrating Pathway (TRIP, Oki and Sud, 1998) developed for use in the UM by Falloon et al. (2007) and implemented in JULES 2.1 by Dr Douglas Clark, Centre for Ecology and Hydrology, was introduced to address these limitations by allowing horizontal water flow via a river network between gridboxes. Each contributing gridbox is given a sequence and a flow direction (see Figure 4.4) to determine which neighbouring cell water travels to. Once the water has entered the channel it can only be stored or moved to the next cell. There is no flooding parameterization.

Conservation of river storage, $S$, over time, $t$, is given by

$$\frac{dS}{dt} = D_{in} - D_{out}$$

where $D_{in}$ is the sum of the inflow from neighbouring gridboxes and $D_{out} = cS$ for

$$c = \frac{u}{dr}$$

where $u$ is the effective flow speed, $d$ is the distance across the gridbox and $r$ is the meandering ratio, given as 1.4 following Oki and Sud (1998). Sensitivity tests were performed (not shown) and, as JULES showed little sensitivity to $u$ over the range 0.5 to 3 m s$^{-1}$ in the TKB and none in the smaller Abisko basin, the default value of 1.5 m s$^{-1}$ was used. The change in channel storage for a gridbox over a given timestep (daily in this study) is then given by

$$S(t_0 + dt) = S(t_0) e^{-cdt} + \left(1 - e^{-cdt}\right) \frac{D_{in}}{c}$$
The map of the river routing network in the Torne and Kalix river basins shown in Figure 4.4 was produced by Bowling et al. (2003a) as part of the PILPS 2(e) experiments from 1/4° maps and digitized data following Lohmann et al. (1996). Figure 4.5 shows two 10-year average hydrographs for the TKB: one obtained by aggregating catchment runoff and the other by running TRIP. As can be seen for a basin the size of TKB, the lag between runoff production in headwaters and streamflow at the outlet point is small (< 2 days) and therefore TRIP makes little change to the timing of flow. However, the channel routing results in a much smoother hydrograph which is more like the observed flow and, as a consequence, all the subsequent JULES runs within the TKB are performed with TRIP.

The main focus in this chapter is to investigate the effect of shrub-tundra on the hydrology in the TKB. Therefore, the changes to the models proposed here will be evaluated both in Abisko and in the TKB. However, the river routing network provided by Bowling et al. (2003a) for Pilps2(e) is appropriate for use in the TKB but not within Abisko. As can be seen in Figure 4.4, where the mask of the Abisko basin is marked in red, flow from gridboxes north of the Abisko catchment contribute to the river discharge at the outlet (68°375’N, 18°875’E) whereas the three boxes in the lower latitudinal line (68°125’N), which do belong to the catchment, do not. As a consequence, TRIP cannot be used in Abisko and instead modelled runoff will be aggregated and scaled with the basin area for comparison with observed discharge at the outlet point.
Figure 4.4: River flow directions (arrows) for the Torne and Kalix rivers with the Abisko basin masked in the red boxes.

Figure 4.5: 10-year average (1989-1999) of measured streamflow (dots) against modelled aggregated runoff (red line) and routed streamflow (black line) over the Torne and Kalix rivers basin.
4.4 Evaluation of the models

Three statistical measures were used to evaluate the different versions of the model proposed in this chapter: bias, root mean square error and the Nash-Sutcliffe efficiency (Nash and Sutcliffe, 1970). The statistics were calculated using the daily time series.

The model bias is given by

\[ \text{bias} = (\bar{M} - \bar{O}) \]  \hspace{1cm} (4.4)

where \( \bar{M} \) and \( \bar{O} \) are the mean modelled and observed discharge respectively. The bias gives an indication as to whether the model over- or under-estimates output compared to observation.

The rmse, given by

\[ \text{rmse} = \sqrt{\frac{\sum_{i=1}^{N} (M_i - O_i)^2}{n}} \]  \hspace{1cm} (4.5)

and where the sum is over \( N \) times, is in the same unit as the data and gives an indication of the “typical” error value.

Efficiency, given by

\[ E = 1 - \frac{\sum_{i=1}^{n} (M_i - O_i)^2}{\sum_{i=1}^{n} (O_i - \bar{O})^2} \]  \hspace{1cm} (4.6)

ranges from \(-\infty\) to 1 where 1 indicates a perfect match between model and observations, 0 to 1 indicates that the model prediction is more accurate than the mean of the data observed and negative values indicate that using observed mean values is a better predictor than using the model.

The results of all the statistical analyses that were performed to evaluate different versions of the models are shown in Tables 4.1 and 4.2 (Section 4.9).

4.5 River flow in JULES

As part of PILPS2(e), Nijssen et al. (2003) found that the parameterization of partitioning between surface and sub-surface runoff determined the ability of models to accurately represent the timing of the snowmelt-induced runoff peak. Models that classified their runoff mostly as surface runoff overestimated the runoff peak and underestimated the recession period.
The runoff formulation as it currently stands in JULES (subsequently referred to as JULES-Ctl) is the same as in the version of MOSES used in PILPS2(e). As can be seen in Figure 4.5 for TKB and Figure 4.6 for Abisko, the hydrographs conform to the conclusions of Nijssen et al. (2003). Modelled peak flow is mistimed such that by the time modelled discharge has started to recede observed discharge is only reaching peak flow. The majority of the snowmelt becomes surface runoff (the surface:sub-surface runoff ratio or SSRR is 6:1 in the TKB and 12:1 in Abisko) because the top soil layer is saturated during the snow season which means that the layers below do not get replenished. As a consequence, at the end of the snowmelt season the deeper soil layers are relatively dry and instead of producing rainfall-runoff as in the observations, JULES-Ctl uses rainfall to replenish the soil column.

Although the efficiency coefficient in Abisko is positive (0.22), unlike in the TKB (-0.30), these results give little confidence in the ability of the model to represent runoff well at high latitudes. As was discussed in Chapter 3, JULES-Ctl considers the Clapp and Hornberger relations in terms of unfrozen water rather than in total (i.e. frozen and liquid) water content (Equations 3.22 and 3.23) which has the effect of substantially limiting infiltration in partly frozen soils and
prevents infiltration into completely frozen soil. While the application of these dependencies may be appropriate for small scales and in the absence of macro-pores, it fails to allow for the heterogeneous spatial distribution of soil moisture in the landscape which depends upon sub-grid processes such as local soil depth, soil properties, vegetation type, wind patterns affecting snow deposition and topography (Seyfried, 1998; Grant et al., 2004; Sayers, 2006).

In the next two sections alternate formulations of soil processes are proposed, which aim to implicitly represent some of these properties, leading to better representation of the timing of runoff events.

4.6 Downward routing of soil moisture in excess of saturation

4.6.1 Description of the scheme

As was discussed in Section 3.6, in JULES soil moisture in excess of saturation for a given layer is passed to the layer above. If the top soil layer is saturated, any excess water becomes surface runoff.

In MOSES 1, supersaturation was avoided by passing the water in excess of saturation to the layer below, i.e. in the opposite direction to that used in JULES 2.1. This formulation had been adopted partly to represent the subgrid variability of permafrost and frozen ground where fractured ice and discontinuous frozen ground can favour water infiltration in the soil within the scale covered by a gridbox (Essery et al., 2001). The scheme was later abandoned as it was found to lead to poor runoff simulations and excessive soil moisture. However, as was seen in Section 3.7, the implementation of the multiple snow layer scheme in JULES 2.1 has lead to a better and warmer representation of subnivean soil temperatures and, by extension, to changes in the partitioning between liquid and frozen soil moisture content. Following these improvements, the scheme used in MOSES 1 is reintroduced to the model to assess if the downward routing may be more applicable with the new multi-layer snow model in permafrost areas. This version of JULES is hereafter referred to as JULES-SD (for “Soil Down routing”)

4.6.2 Evaluation of JULES-SD

In TKB (Figure 4.7-a), JULES-SD overestimates maximum flow rates by less than JULES-Ctl, which leads to a smaller rmse (Table 4.1). Whereas peak flow in JULES-Ctl ends when observed peak flow starts, in JULES-SD it starts when
Figure 4.7: 10-year average (1989-1999) observed streamflow against (a) model streamflow in the Torne and Kalix rivers basin and (b) model runoff in the Abisko catchment with JULES-Ctl and JULES-SD.
observed peak flow ends. This is due to the fact that the water diagnosed as being in excess of saturation first has to travel through the whole soil column, replenishing each layer if necessary along the way, to finally come out only when the bottom layer is wet enough to generate significant subsurface flow (SSRR = 1:9). As a consequence the summer rainfall-induced streamflow period is reproduced well as, unlike in JULES-Ctl, the soil has not dried out by the end of the snowmelt season and Dunne overland flow can occur during rainfall events. Although this formulation raises the efficiency coefficient of the model to 0.34 in the TKB, the performance of the model is even better at Abisko where efficiency is 0.48. The first effect on runoff that “pushing” the saturation excess water down the soil column has is to smooth the hydrograph. The spikiness of the runoff time series in JULES-Ctl is due to snowmelt being immediately translated as surface runoff because the top soil layer is saturated. In JULES-SD this water is forced to the layers below where some is kept in the soil column as storage, thus removing the high frequency variability from the snowmelt-induced runoff.

In the absence of extensive snow depths measurements in the TKB, the delay between snowmelt and snowmelt-induced streamflow production cannot be assessed. Nevertheless, it is highly likely that JULES-SD overestimates this delay. One possible contributing factor in this lateness is that JULES assumes uniform deep soils of 3 m, whereas soils in TKB are much thinner on average because of exposed bedrock in mountainous regions and permafrost (Lyon et al., 2009). JULES-SD is essentially a simple modification to soil parameterization which uses little physical reasoning. In the next section a more complex representation of soil processes for use in heterogeneous Arctic landscapes is proposed.

4.7 Allowing infiltration in partially frozen soils

4.7.1 Description of the scheme

Niu and Yang (2006) implemented a formulation of the hydraulic conductivity (hereafter referred to as the NY scheme) in the Community Land Model version 2.0 (CLM2.0) where a permeable fraction within the ice fraction is introduced in order to represent the permeability of frozen soils and the subgrid variability of the Arctic landscape in large scale modelling.

The water flux within a gridbox is expressed as

\[ W = (1 - F_{frz})W_u + F_{frz}W_f, \]  

(4.7)
where \( W_u \) and \( W_f \) are the water fluxes in the unfrozen and frozen fractions respectively and where \( F_{frz} \), the fractional impermeable frozen area, is given by

\[
F_{frz} = e^{-a(1-\theta_f/\theta_{sat})} - e^{-a};
\]

(4.8)

where \( \theta_f \) is the volumetric frozen water content and \( a \) is a scale-dependent parameter taken as 3 following Niu and Yang (2006). The hydraulic conductivity is given as

\[
K = (1 - F_{frz})K_{sat} \left( \frac{\theta}{\theta_{sat}} \right)^{2b+3}.
\]

(4.9)

Two criteria distinguish Equation 4.9 from the formulation of \( K \) in JULES in Equation 3.23:

1. \( K \) is only considered as a function of the unfrozen fraction in Equation 3.23 rather than as a function of the total soil moisture content as in Equation 4.9, allowing for water movement through frozen soil.
2. \( K_{sat} \) is scaled down by the impermeable fraction and thus reduces the value of \( K \).

Figure 4.8 shows the relative hydraulic conductivity \( (K/K_{sat}) \) for JULES and the NY scheme for \( \theta = \theta_{sat} \) and for different values of \( a \). In JULES, \( K/K_{sat} \) decreases rapidly from 1 for \( \Theta_f = 0 \) to close to 0 for \( \Theta_f > 0.3 \). On the other hand, the NY scheme allows for more water to travel in the soil column and to be stored in the landscape by permitting water flow within the ice fraction. The empirical parameter \( a \), used in Equation 4.8, controls the amplitude of the water flow through the frozen fraction. The greater \( a \) is, the sharper the transition is from a high to a low \( K/K_{sat} \). Niu and Yang (2006) found that this formulation of \( K \) resulted in improved river discharge simulations in CLM2.0 by providing longer recession periods and smaller river flow peaks during the spring snowmelt.

Figure 4.9 shows the relationship between the ice fraction and the permeable area in Equation 4.8 for four different values of \( a \). As JULES does not explicitly calculate a permeable area, the ice fraction is assumed to be completely impervious (black line). By including \( F_{frz} \), the NY scheme allows for a permeable fraction within the frozen soil. The permeability of \( F_{frz} \) is determined by \( a \) such that the higher it is, the sharper the transition is from permeable to impermeable.

The implementation of the NY scheme in JULES is hereafter referred to as JULES-NY.
Figure 4.8: Relative hydraulic conductivity (with $b = 5.0171$) as a function of the frozen water content fraction ($\Theta_f$) for different $a$.

Figure 4.9: Fractional permeable area as a function of the ice fraction for different $a$. 

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4.7.2 Evaluation of JULES-NY

Flow rates with JULES-NY are very similar to those with JULES-SD as Figure 4.10 and the surface:subsurface runoff ratio, rmse, bias and efficiency in Table 4.1 and 4.2 show. Both schemes allow more water in the soil column than JULES-Ctl and, as a consequence, both are dominated by subsurface flow.

Although the runoff and streamflow patterns between JULES-SD and JULES-NY are similar, the process through which water travels through the soil column differs. In JULES-SD the shallow top soil layer quickly gets saturated in autumn when the layer freezes and any additional water input to the layer is therefore “pushed” to the layer below. This process occurs until all the layers are saturated, at which point there is enough water in the bottom layer to generate substantial subsurface runoff. On the other hand, because JULES-NY allows infiltration through partly frozen soil, the soil never gets saturated. The water moves steadily in the column throughout the season and subsurface flow starts earlier than with JULES-SD in the TKB.

The smoothness of the hydrograph and the simulation of rainfall-induced discharge result from the same processes as those explained in Section 4.6.2 and therefore are not repeated here.

4.8 JULES with TOPMODEL

As was seen in Section 3.6, runoff can occur in three ways: as saturation or infiltration overland flow or as slow response sub-surface flow. The number of responses possible following a single rainfall or snowmelt event within a basin imply a nonlinear relationship between the event and its responses. Kirkby and Weyman (1974) were the first to develop a simple model where the runoff-response from a rainfall event was calculated using relationships between elevation, slope angle, soil moisture availability and runoff. Since then models derived from the TOPography-based hydrological MODEL (TOPMODEL) have used these relationships for a number of applications (e.g Beven and Kirkby, 1979; Sivapalan et al., 1987; Quinn et al., 1995; Beven, 1997; Stieglietz et al., 1997; Gedney and Cox, 2000; Clark and Gedney, 2008). The description of the model that follows is adapted from these papers.

4.8.1 Description of TOPMODEL

Three assumptions are at the core of every TOPMODEL-based model:
Figure 4.10: 10-year average (1989-1999) measured streamflow against model (JULES-Ctl, JULES-SD and JULES-NY) (a) streamflow in the TKB (b) runoff in Abisko.
1. The local downslope flow, \( q_l \), is balanced by a spatially uniform steady recharge rate, \( R_r \), from a local upslope area, \( \alpha_l \), such that

\[
q_l = \alpha_l R_r \tag{4.10}
\]

2. The saturated hydraulic conductivity declines exponentially with depth as follows:

\[
K_{sat}(z) = K_{sat}(0)e^{-fz} \tag{4.11}
\]

where \( z \) is depth into the soil profile (positive downward), \( K_{sat}(0) \) is the saturated hydraulic conductivity at the surface and \( f \) is the saturated hydraulic conductivity decay factor, taken to be 3 in this version of the model.

3. The water table is parallel to the soil surface such that the downslope flow can be related to the local slope angle, \( \beta \), as

\[
q_l = T(z_{wl}) \tan \beta \tag{4.12}
\]

where \( T(z_{wl}) \) is the local transmissivity gradient given by integrating Equation 4.11 vertically from the local water table depth, \( z_{wl} \), to the bottom of the profile, \( z_{wmax} \)

\[
T(z_{wl}) = \int_{z_{wl}}^{z_{wmax}} K_{sat}(z)dz \tag{4.13}
\]

These three assumptions lead to an analytical solution (shown in full in Sivapalan et al., 1987) for the local water table depth using the relationship between the mean and the local water table depths within the catchment such that

\[
z_{wl} = z_w - \frac{1}{f}(\chi - \bar{\chi}) \tag{4.14}
\]

where \( z_w \) is the mean water table depth and \( \chi \) and \( \bar{\chi} \) are the local and mean topographic indices respectively.

The topographic index, which is used as an index of hydrological similarity where all points with the same value are assumed to have the same hydrological response (Beven, 1997), is calculated as

\[
\chi = \ln \left( \frac{\alpha}{T_0 \tan \beta} \right) \tag{4.15}
\]

where \( \alpha \) is the upslope area that drains through a point and \( T_0 \) is transmissivity when the water table is at the soil surface. Information on the topography of
the basin, obtained from a digital terrain model, is required to calculate the
distribution of $\chi$ in the catchment. To calculate the fraction of the catchment
that is saturated at the surface, $f_{sat}$, $z_{wl}$ is set to 0 in Equation 4.14. Saturation
occurs where $\chi > \bar{x} + f_{zl}$ and $f_{sat}$ is obtained by integrating the probability
distribution function of the topographic index up to this value.

Integrating along the watershed channel network gives the baseflow as

$$Q_{base} = \frac{K_{sat}(0)}{f} e^{-\bar{x}} e^{-f_{zl}}$$  \hspace{1cm} (4.16)

4.8.2 Extension of TOPMODEL in JULES

JULES 2.1 includes a TOPMODEL-based scheme (JULES-TOP) following Ged-
ney and Cox (2000). The scheme is an extension of the models introduced in
Sivapalan et al. (1987) and Stieglietz et al. (1997) who adapted TOPMODEL for
use in LSMs. The changes to the physics introduced in Section 4.8.1 are

1. A deep soil layer (9 m thickness) is added to the four standard JULES soil
layers and is used to track $z_{wl}$ when it falls below the standard four soil
layers.
2. Assumption (2) in Section 4.8.1 is relaxed and $K_{sat}$ is only assumed to
decline exponentially in the saturated zone in the added deep soil layer.
3. JULES assumes that the local area within the grid box where the local $\chi$
equals that of the gridbox mean corresponds to areas of gridbox mean water
table depth (Gedney and Cox, 2000). It then uses the mean and standard
deviation of the topographic index in each gridbox derived from higher
resolution data, to describe a 2-parameter gamma probability distribution
function for $\chi$. However as this method is too computationally expensive
to integrate, JULES 2.1 then fits an exponential relationship between $\chi_{crit}$
and $f_{sat}$.

The relationship between the maximum baseflow, $Q_{base}^{max}$, and the actual baseflow
then allows the calculation of the critical $\chi$ value when the water table comes to
the surface such that

$$\chi_{crit} = \ln \left( \frac{Q_{base}^{max}}{Q_{base}} \right)$$  \hspace{1cm} (4.17)

$\chi_{crit}$ relates the distribution function of $\chi$ to the saturated fraction within the
gridbox as all points with $\chi > \chi_{crit}$ are assumed to be at saturation and Dunne
overland flow will be generated at the rate of the product of $f_{\text{sat}}$ and the water input to the surface (rainfall and snowmelt).

This version of the model is hereafter referred to as JULES-TOP.

4.8.3 Evaluation of JULES-TOP

There is little difference between the JULES-Ctl and the JULES-TOP hydrographs in Figure 4.11, although the efficiency coefficients suggest that the implementation of a TOPMODEL-based scheme in JULES produces poorer streamflow rates. The high SSRR (80:1 in TKB and 29:1 in Abisko) shows that JULES-TOP is even more dominated by surface runoff and manages to produce even less subsurface and base flow than JULES-Ctl. Both schemes cause the top soil layer to saturate early in the snowmelt season such that infiltration of the snowmelt is subsequently impaired. As a consequence, snowmelt produces surface runoff rather than soil moisture. However when the soil falls below saturation JULES-Ctl switches to only producing runoff via subsurface flow or infiltration excess surface flow. In contrast, JULES-TOP also generates surface runoff, as a result of $f_{\text{sat}} > 0$ where the local water table comes to the surface.

As the version of CLM2.0 used by Niu and Yang (2006) included a simple TOPMODEL-based runoff scheme, JULES-TOP was also tested with the NY scheme. This version is hereafter referred to as JULES-TOP-NY.

The results between the TKB and Abisko are mixed. In the TKB, this version of the model has the highest efficiency coefficient and smallest rmse and bias. JULES-TOP-NY is able to capture efficiently (1) the timing of the onset of the snowmelt-induced river flow, (2) the duration of peak flow (3), the streamflow rates during peak times and (4) the beginning of the recession period (Figure 4.11). Flow rates outside the snowmelt season are still underestimated, although the rainfall-induced river flow events are closer to observations than with the other models. However, streamflow when snow is on the ground (∼ November to April) is the lowest of all versions.

Unlike with JULES-NY, most of the streamflow with JULES-TOP-NY is produced by surface runoff. As was seen previously, the NY scheme allows more water through the soil column than JULES-Ctl and JULES-TOP. Added to JULES-TOP, this feature leads to delayed streamflow, smaller peak flow rates and higher flows during the recession period, compared to JULES-TOP and JULES-Ctl, that closely match observations. This improvement with JULES-TOP-NY is due to differences between the NY scheme and JULES-TOP in the bottom boundary condition. In JULES-TOP-NY, if the added deep layer is very wet and not able
Figure 4.11: 10-year (1989-1999) catchment average observed river flow against (a) modelled (JULES-Ctl, JULES-TOP and JULES-TOP-NY) streamflow in the TKB (b) modelled runoff in Abisko.
to generate baseflow as fast as the NY scheme delivers water through the soil column, the water table rises, potentially saturating the soil. This switch from infiltration into the soil to increased surface runoff production, which occurs at the heart of the snowmelt season, produces a better match to observed streamflow, as the highest efficiency and the lowest rmse and bias of all runs within the TKB corroborate. The springtime saturation also leads to a better simulation of rainfall-induced river from mid-July to November because the soil layer is still wet and so can still generate surface runoff.

While this model set-up appears to be the most appropriate for representing peak flows in nival hydrological regimes over the TKB, both JULES-TOP and JULES-TOP-NY make little difference to the runoff in the smaller Abisko catchment. Potential reasons behind the poorer performance of the model in Abisko, also applicable to the other model set-ups, are discussed below.

### 4.9 Potential sources of error

![Table 4.1: Rmse, bias and efficiency of modelled vs observed streamflow at the mouth of the Torne and Kalix rivers, and modelled surface:sub-surface runoff ratio (SSRR) over the whole catchment.](image)

![Table 4.2: Rmse, bias, efficiency of modelled vs observed runoff and modelled surface:sub-surface runoff ratio (SSRR) in the Abisko catchment.](image)

Tables 4.1 and 4.2 summarise the results of each scheme. Figure 4.12 shows modelled and observed streamflow (TKB) and runoff (Abisko) rates for all the schemes tested in this chapter, so as to make comparisons between schemes easier.
Figure 4.12: 10-year (1989-1999) catchment average observed river flow against modelled (a) streamflow in the TKB and (b) modelled runoff in Abisko for all parameterizations studied.
Firstly, in order to reduce errors from streamflow simulations, the new parameters introduced in JULES-NY, JULES-TOP and JULES-TOP-NY could be calibrated. Calibration involves adjusting parameter values so that the model more closely matches the behaviour of the real system it represents (Gupta et al., 1998). Many methods adapted to the increasing complexity of hydrological models have been devised but most have in common the automated computation of a very large number of values being tested to find the optimal value of a single or multiple parameters (e.g. Spear and Hornberger, 1980; Beven and Binley, 1992; Duan et al., 1992; Gupta et al., 1998). Increasing the complexity of a model requires the addition of extra parameters which, if they do not come with additional observational data for validation, add more uncertainties to model results (Beven, 2006). In addition, complex land surface models like JULES can make it difficult to diagnose the exact sources of model errors because the results may be net effects of possibly conflicting signals. Equally, the same results might arise from quite different sets of processes, which is known as the concept of “equifinality” (Beven, 1993). Parameters are often conceptual representations of abstract basin characteristics and thus can often require very different values for different basins. Yet parameter transferability between basins is essential because of the limited number of gauged basins, particularly in the Arctic, which makes streamflow predictions in ungauged basins a major challenge to the hydrological modelling community (Sivapalan et al., 2003). Calibration was not performed in this study because this work aims to improve the general representation of snow and runoff processes at high latitudes, rather than to tune the model to represent a particular catchment. As a result, some of the default parameters values used in JULES may not be appropriate for use in the two basins. For example, of significance in the mountainous Abisko catchment, the soil profile is 3-m deep in all but JULES-TOP and JULES-TOP-NY where it is 12-m deep. Mountains are likely to have thinner or no soil profiles, leading to surface runoff-dominated hydrographs. Secondly, the resolution at which JULES is run may not allow it to capture a range of important sub-grid processes. For example, the single model elevation in a 1/4° gridbox is not representative of the range of elevations found in reality. This is particularly significant in the Abisko catchment, which is situated within a mountainous area. Equally, the topography of the basin (e.g. slope and aspect) is not represented. As a consequence, the modelled snowmelt rates are uniform where actual rates are likely to vary considerably because of topography and altitudinal gradients. Both Wiltshire (2006) and Parajka et al. (2010) identified the implementation of a sub-grid representation for topographic heterogeneity as be-
ing a critical model development for the performance of JULES over mountainous regions.

Thirdly, Bowling et al. (2003a) noted that, for some LSMs that participated in the PILPS2(e) experiment, the hydrology of the small basins situated above the treeline in the TKB was problematic because some models (including MOSES) failed to represent the attenuation of high flows by the absorption capacity of peatland and bogs which dominate the basins.

Fourthly, the vegetation in the Abisko catchment is dominated by open and closed shrubland, unlike most of the TKB which is woodland and wooded grassland. While a number of functional type-specific parameters define the characteristics of different vegetation covers, some of the model parameterizations are more appropriate for dense tall vegetation types than sparse tundra types. For example, the effect of the vegetation cover on the calculation of the tile albedo is represented solely by the roughness length, and by extension the height, of the surface type (see Equations 3.13 and 3.15). These formulations do not take into consideration the fact that tall shrubs can get buried under the snowpack in the winter (Pomeroy et al., 2006; Marsh et al., 2010). Furthermore, the lack of horizontal interaction between tiles precludes the model from representing processes that are critical in open tundra landscapes such as the advection of heat from snow-free patches to the snowpack or wind-borne snow transport from open to closed shrub canopies.

Finally, there may be other physical differences between Abisko and the TKB that were not identified in the previous points and which would lead to improved model simulations in the latter than in the former.

4.10 Conclusion

In this chapter, the land surface model JULES was evaluated against snow cover and streamflow in two Arctic basins. Sources of errors in model streamflow were attributed to the partitioning between surface and subsurface runoff. In order to address this issue, three alternative formulations of soil processes were proposed. Combining the two formulations that include fractional covers (saturated fraction and permeable fraction) to represent sub-grid heterogeneity considerably improved model performance at one of the sites. Possible sources of errors for the little difference any of the formulations made at the shrub-tundra site were discussed.

Fractional sub-grid coverage is often used in models to represent landscape heterogeneity. For example, in this study JULES uses a snow depth-dependent snow
cover fraction to account for patchiness, the NY scheme introduces a permeable fraction in the frozen soil to allow more water through the soil column and TOPMODEL calculates surface runoff as a function of a saturated fraction. Other models use a fixed vegetation fraction, $F_v$, to distinguish snow-covered from non-snow-covered areas within tiles (e.g. Verseghy et al., 1993). Improving the representation of all processes in mountainous regions is beyond the scope of this study. On the other hand, addressing the representation of shrub characteristics is one of its objectives. As was discussed in Section 1.6, some shrubs bend and get buried in the snowpack. Therefore, omitting this process in models is likely to cause lower albedo values than in reality and affect the modelled energy balance over shrub-tundra sites. In the next chapter, a high resolution shrub bending model is developed to establish the vegetation fraction over a snow season cycle at a shrub-tundra site. A $F_v$ parameterization is then derived from the model in an attempt to provide a better representation of runoff processes in the Abisko basin in particular and of snow-shrub interactions in general.
Chapter 5

The Shrub Bending Model (SBM)
5.1 Background

The study of shrub bending mechanisms belongs to the science of plant biomechanics. However, this phenomenon has raised much interest amongst high latitude hydrologists and micrometeorologists over recent years (e.g. Sturm et al., 2005b; Pomeroy et al., 2006; Marsh et al., 2010) as the aspect of shrubs (i.e. whether they are erect, bent or buried) affects the water and energy balances in a number of ways.

A single snowfall event can transform the areal snow cover from 0.1 to 1.0 million km\(^2\) in a matter of days (Walsh et al., 2004), leading to an increase in albedo over a very short timescale. In springtime, albedo starts to decrease before the shrub stems are exposed because of the penetration of shortwave radiation into the snowpack (Warren, 1982; Baker et al., 1991; Hardy et al., 1998). Once exposed, branches can be up to 20\(^\circ\)C warmer than the surrounding snow and, by redistributing energy as longwave radiation and advective sensible heat, enhance the snow melt rates (Pomeroy et al., 2006; Strack et al., 2007; Essery et al., 2008). This is illustrated in the sensible heat flux measurements in Figure 5.1 taken at two sites within the WCRB: the low, P2, and the tall, GB2, shrub sites situated about 300 m away from one another (see Section 2.2.1 for a description of the two sites). Heat fluxes are very different before day 121 but are similar after. There is little \(H\) exchange between the atmosphere and the surface before day 121 at P2 because the shrubs are buried in the snowpack, whose surface temperature is close to the air temperature. On the other hand, the temperature difference between the exposed branches at GB2 and the atmosphere causes positive \(H\) from the surface to the atmosphere, even though there is still snow on the ground. As a consequence, energy exchanges at a shrub-tundra landscape where tall shrubs are bent under the snow will be similar to those at a low shrub site.

Schmidt and Gluns (1991) found that the snow interception process in needleleaf trees occurs through the trapping of snow crystals on and between the needles and the branch stem. Snow accumulates through the cohesion of new snow crystals until bridging between needles, then between smaller branches, occurs. Field observations by the author of the present study suggest that snow interception by birch shrubs follows a similar pattern in which snow crystals are first collected where small stems intersect (Figure 5.2a-b). More in depth field studies are required to confirm this hypothesis. Willows have an additional way of trapping snow: they keep their marcescent (i.e. withered) leaves through the winter which provides a large surface area for snow interception (Figure 5.2-c).
Sturm et al. (2005b) and Pomeroy et al. (2006) suggested that bending occurs at near freezing temperatures during storms associated with strong winds and heavy snowfall. Following field observations of an overnight snowfall that led to partial bending of shrubs in the WCRB, the author believes that a storm event would, on the contrary, blow snow off branches. The gentle snowfall witnessed by the author, on 30 April 2008, allowed snowflakes sufficient time to bond and presented no disturbance to the adhesion of snow on branches, which caused the willow in Figure 5.3 to bend overnight. In addition, if wind speed caused bending, it is likely that there would be preferential bending amongst the more exposed shrubs, for example on the edge of shrub patches. Such a distribution was not observed in the WCRB and agrees with findings from Beismann et al. (2000) who found that out of 8 species of willows tested, the only one regularly subjected to snow loads (Salix appendiculata) was the only one that showed no consistency, structural or geographical, in elasticity modulus variation. This behaviour is consistent with the apparent randomness of the distribution of bent shrubs in the WCRB. Further field observations are therefore needed to gain a better understanding of bending behaviour and to determine the response of individual willows to different meteorological events.

Schmidt and Pomeroy (1990) and Schmidt and Gluns (1991) found that air temperatures between \(-3^\circ C\) and \(0^\circ C\) promote interception through (i) decreased snow density (ii) increased cohesive (liquid-like) properties of snow (iii) decreased
Figure 5.2: Intercepted snow on birch shrubs (a and b) and willows (c).
Figure 5.3: Willow shrub before (a) and after (b) a single snowfall.
crystal rebound by which crystals either rebound off the branch and / or erode snow already intercepted. With an average of 0.7°C, the overnight snowfall of the 30 April 2008 does substantiate the suggestion that bending occurs at near freezing temperature.

Beismann et al. (2000) also found that, when subjected to the same load as the other tested species, *S. appendiculata* was the least brittle and branches several centimetres thick would bend rather than break. By extrapolating these results outside the context of willows thriving in avalanche-prone regions, it seems probable that other *Salix spp.* subjected to similar strains would have developed a similar mechanism to sustain heavy snowload. In addition to preventing breakage, adaptation through bending could provide protection against abrasion by blowing snow grains, desiccation and low winter air temperatures. However, the ecological significance of shrub bending warrants further investigation that is beyond the scope of this study.

In this chapter, a simple model that reproduces the bending processes is presented. A description of the parameters and of potential sources of errors follows the description of the physics of the model. The model is then evaluated against field measurements and a parameterization of the exposed shrub fraction, derived from the model, for use in energy balance models is proposed. Finally, the sensitivity of the model to study-specific parameters is assessed.

### 5.2 Description of the Shrub Bending Model

The Shrub Bending Model (SBM) was built on the principle that branches of shrubs can be modelled by engineering analogy with the elastic theory (or the Elastica) of the large deformation of cantilever columns for which all the basic equations presented here can be found in structural engineering textbooks (*e.g.* Marshall et al., 1990; Megson, 1996).

#### 5.2.1 Model physics

Referring to Figure 5.4, the curvature of a column can be expressed as $d\theta/ds$ where $\theta$ is the angle between the branch and the vertical and $ds$ is the length of a segment along the branch. The bending moment in the horizontal $x$ direction imposed on the branch is related to the curvature such that

$$
\frac{d\theta}{ds} = \frac{Mg}{EI}x
$$

(5.1)
Figure 5.4: Schematic representation of the bending of a cantilever column
where $M$ is the load applied on the branch, $E$ is the Young’s (or elastic) Modulus and $I$ is the second moment of area about the axis of bending defined as

$$I = \frac{\pi r^4}{4} \tag{5.2}$$

where $r$ is the radius of the branch in cross section. This equation indicates that the curvature of the branch is not only related to the load on the branch but also inversely proportional to the flexural rigidity, which is a product of the shape of the branch $(I)$ and its material properties $(E)$. Timoshenko and Gere (1961) recognised that Equation 5.1 has a dynamic analogy with the equation governing the oscillatory motion of a pendulum such that

$$L = \frac{1}{2k} \int_0^\alpha \frac{d\theta}{(\sin^2 \frac{\theta}{2} - \sin^2 \frac{\phi}{2})^{1/2}} \tag{5.3}$$

where

$$k = \left(\frac{Mg}{EI}\right)^{1/2} \tag{5.4}$$

and $L$ is the length of the branch, $\alpha$ is the deflection angle at the upper, free end of the branch and $\phi$ is a parameter introduced to simplify the integral in equation 5.3 such that

$$\sin \frac{\theta}{2} = p \sin \phi \tag{5.5}$$

where

$$p = \sin \frac{\alpha}{2} \tag{5.6}$$

Differentiating Equations 5.5 and 5.6 gives

$$d\theta = \frac{2p \cos \phi d\phi}{\cos(\theta/2)} = \frac{2p \cos \phi d\phi}{(1 - p^2 \sin^2 \phi)^{1/2}} \tag{5.7}$$

Combining Equations 5.3 and 5.7 gives

$$L = \frac{1}{k} \left[ F(p) \right] \tag{5.8}$$

Niklas and O’Rourke (1987) recognised that the integral in Equation 5.8 had the form of a complete elliptic integral of the first kind, designated here as $F(p)$. Combining Equation 5.4 with Equation 5.8 gives

$$Mg = F(p)^2 \frac{EI}{L^2} \tag{5.9}$$
Equation 5.3 to Equation 5.9 were taken from Niklas and O’Rourke (1987). This relationship is appropriate to calculate the bending of columns that are vertical before a load is applied. However, primary branches of shrubs are usually inclined at some angle \( \phi_0 \), which therefore necessitates a reinterpretation of Equation 5.8 such that

\[
kL = F(p, \pi/2) - F(p, \phi_0)
\]

(5.10)

where the elliptic integral of the first kind on the right of the equation is complete (an elliptic integral is said to be complete when the amplitude of the angle is \( \pi/2 \)) and the second is incomplete.

As well as calculating the shape of the branch, the model needs to be able to inform an energy balance model of the position of the branches with regards to the atmosphere i.e. whether they are buried under or protruding above the snowpack. Hence, SBM needs to calculate the vertical and horizontal deflections, \( z \) and \( x \) respectively, of a branch with regards to the shape of its curvature.

The coordinates along the branch can be calculated by a parametric curve, following Aristizábal-Ochoa (2004), such that

\[
x(\phi) = \frac{2p}{k}(\cos \phi_0 - \cos \phi)
\]

(5.11)

and

\[
z(\phi) = \frac{1}{k}[2E(p, \phi) - 2E(p, \phi_0) - F(p, \phi) + F(p, \phi_0)]
\]

(5.12)

where \( E(p, \phi_n) \) is an incomplete integral of the second kind and \( F(p, \phi_n) \) of the first kind. Points along the branch are then simply considered as exposed if \( z(\phi) > \) snow depth or buried if snow depth \( > z(\phi) \).

Pomeroy et al. (2006) defined shrub spring-up to be a “mechanical process which is a function of snow depth and of snow grain bond weakening rate due to wet snow metamorphism and hence occurs branch by branch and shrub by shrub”. However, although the mechanism behind shrub spring-up is understood, no empirical data relating the snow properties to the occurrence of spring-up are, to the author’s knowledge, available. As a consequence, the spring-up parameterization was approximated to what was considered to be a plausible explanation of the process. Therefore, following visual field observations, spring-up and snow unloading was parameterized as occurring when snow depth is less than the height half way between the tip of the branch and its highest point.
5.2.2 Parameters

The parameters needed to implement the structure of the shrub, branch height and primary branch radius at breast height (1.30 m), were established from field measurements during site visits in Spring 2008 and 2009. Random numbers with specified means and standard deviations were generated to constitute the girth and height of the branches.

5.2.2.1 Elastic modulus

The value used for the elastic (or Young’s) modulus of wood, $E$, was taken from Johnson (1987) who determined the $E$ of *Salix glauca*, a species of willow found in WCRB, to be $4.197\times10^{10}$ N m$^{-2}$ ± 2183. $E$ was parametrised as constant even though its value is known to be affected by (a) temperature (Schmidt and Pomeroy, 1990; Lundström et al., 2008), (b) water availability (Niklas and O’Rourke, 1987) and (c) the direction of the measurements to the grain (Niklas, 1992).

Elasticity measurements conducted within a controlled environment in which samples of branches would be cut and subjected to loading before and after drying, at different temperatures and for bending in different axes would be beneficial to get more representative $E$ values. However, the author believes that the lack of species-specific information available on the relationship between the elasticity of shrubs and the influences of (a), (b), and (c) above may generate model errors that are not significant compared to other potential sources of error (for example, as discussed in Section 5.2.3). Therefore the value from Johnson (1987), obtained using procedures specified by the American Society for Testing and Materials, at -5°C, was deemed appropriate for use in SBM.

5.2.2.2 Snow-free vegetation fraction

In order to calculate the fraction of shrubs that are protruding above the snow-pack, $F_v$, the constant snow-free shrub fraction, $F_{v0}$, was calculated with the information available from the high-resolution digital aerial photograph used in Bewley et al. (2007) and reproduced in Figure 5.5. It shows a 30 m x 30 m grid from which Bewley et al. (2007) calculated $F_v = 0.71$ for average snow depth, approximated from snow surveys, as being $\sim 0.30$ m. $F_v$ was obtained from Figure 5.5 by applying an intensity threshold to the image to produce a binary mask. Using this information and from trials during the development stage of the model, $F_{v0}$, which is a constant, was then found to be 0.75.
5.2.2.3 Plant Area Index, foliage factor and vegetation fraction

The relationship between the vegetation fraction, $F_v$, and the plant area index, $PAI$, defined here as the total projected one-sided plant area per unit surface of ground ($PAI = A_{proj}/A_{ground}$), was introduced by Nilson (1971) and defined as

$$F_v = 1 - e^{-PAI}. \quad (5.13)$$

$A_{proj}$ is calculated as

$$A_{proj} = \sum_{i=1}^{n} (x_i 2r_i) f \quad (5.14)$$

where $n$ is the number of branches and the terms in the bracket are the cumulative products of the diameter and projected length of all the primary branches and $f$ is a foliage factor, introduced to increase the interception surface area because secondary branches and marcescent leaves are not explicitly modelled in SBM. $A_{ground}$ was calculated as the square of the average maximum horizontal projection of branches.

Research on interception of snow in coniferous forests has shown that interception efficiency can be reduced with increasing load (Schmidt and Gluns, 1991). As a result, an effective foliage factor, $f_{eff}$, is calculated, following Hedstrom and Pomeroy (1998), as
\[ f_{\text{eff}} = f(1 - S_{\text{rel}}) \]  
(5.15)

where \( S_{\text{rel}} \) is a relative canopy storage capacity parameter defined as

\[ S_{\text{rel}} = \frac{S}{S_{\text{max}}} \]  
(5.16)

for \( S \) being the intercepted snowfall and \( S_{\text{max}} \) being the maximum canopy storage capacity, defined as the product of the \( PAI \) and a mean species snow interception value, \( \overline{S} \). Schmidt and Gluns (1991) published comprehensive empirical data of \( \overline{S} \) for needleleaf branches which have been used in numerous models (e.g. Hedstrom and Pomeroy, 1998; Pomeroy et al., 1998; Essery, 2002). However, no empirical data of interception capacity are available for shrubs with marcescent leaves and a nominal value of \( S_{\text{max}} = 5PAI \) was used. During melt \( S_{\text{rel}} \) is set to 0 for protruding branches.

An effective \( PAI, PAI_{\text{eff}} \), which represents the total one-sided plant area protruding above the snowpack per unit surface of ground, is calculated as

\[ PAI_{\text{eff}} = A_{\text{proj}} \frac{f_{\text{eff}}}{A_{\text{ground}}} \]  
(5.17)

Following Equation 5.13, the fraction of exposed vegetation at each timestep can then be calculated as

\[ F_v = 1 - e(-PAI_{\text{eff}}) \]  
(5.18)

As the model is driven by snowfall and snowmelt only, there is no parameterization of unloading of intercepted snow, notably by sublimation, wind or melt when temperatures are close to or above freezing point.

### 5.2.3 Potential sources of error

SBM only explicitly models primary branches. Coupling SBM to a structural biological model, such as the L-system - or Lindenmayer system - (Lindenmayer, 1968) which uses a rewriting string system, would allow us to have a more realistic shrub structure but would introduce complications that were not deemed necessary. The purpose of SBM is to ascertain the relationship between the vegetation fraction \( (F_v) \) and snow depth in order to derive a function that can serve as input in an energy balance model, not to model the behaviour of individual branches under the weight of snow, which is nonetheless a means to the end. The
parameter $f$, presented in Section 5.2.2.3, was instead introduced to implicitly simulate the foliage surrounding what is effectively the primary branch. Another assumption used in the model is that of the shape of the branch itself. For simplicity, the branches of the shrub were modelled as untapered cantilever columns. However, tapering is known to cause the top of the branch to bend more than its base i.e. the branch does not bend with uniform curvature (Johnson, 1987). One way to partly overcome this issue is by applying the snow load at the tip of the branch. In reality, snow is spread along the branch but the model does not account for this. By accentuating the weight of the snow at the tip of the branch, the top of the branch bends more than its base. Plasticity or non-recoverable deformations are not modelled, meaning that branches return to their pre-load position and shape after they spring-up. As the standard deformation in biological materials is usually $\sim 0.1\%$ (Niklas, 1992), it was therefore not considered significant enough to include in the model. There have been few field observations of the processes modelled in SBM. More field studies are therefore needed to gain a better understanding of the timescale over which bending occurs, of the distribution of the bent shrubs and of the influence of weather conditions.

5.3 Evaluation of the model performance

5.3.1 Shrub bending and spring-up simulations

Figure 5.6 illustrates a modelled shrub at different bending stages. The sections of branches that are protruding above the snowpack are shown as dark brown and the sections in the snowpack are light brown. The figure only shows the part of the branch that is explicitly modelled ($A_{\text{proj}}$) i.e. there is no representation of the foliage, implicit in the model in $f$. The shrub prior to any load being applied is shown in Figure 5.6a. Figure 5.6b-e show the progression of the shrub’s shape with increasing snowload. All branches bend at different rates because of their individual height and girth. For example, three branches have bent relatively quickly in Figure 5.6 in relationship to the amount of snowfall, illustrated in the shallow snow depth (light brown). When the snow is deeper than the distance between the ground and the tip of the branch, the branch is progressively submerged, the tip and the base getting buried but the middle of the branch still protruding above the snowpack (Figure 5.6c). As more snow falls, more of the shrub gets buried (Figure 5.6d-e). During snowmelt, the branches spring-up faster than they bent and many branches are already erect
Figure 5.6: Phases of the shrub when snow-free (a), during bending and burying at $SWE = 47$ mm (b) $SWE = 95$ mm (c) $SWE = 191$ mm (d) $SWE = 480$ mm (e) and during snowmelt at $SWE = 107$ mm (f). The exposed part of the shrub is in dark brown, the buried part in light brown. See text for details.
while there is still a considerable amount of snow on the ground. This is due to
the hysteretic relationship between the accumulation and the snowmelt cycles,
discussed below. At the end of the run, the shrub regains its initial shape (Figure
5.6a) as plastic deformation is ignored.

### 5.3.2 Model evaluation against measured vegetation fraction

Figure 5.7 shows the modelled relative vegetation fraction ($F_v/F_{v0}$ where $F_{v0}$ is
the snow-free vegetation fraction) as a function of the relative snow depth ($S_d /
average shrub height$) evaluated against three measured $F_v$. The first observed
$F_v$ (0.71) was calculated by Bewley et al. (2007) from Figure 5.5, as explained
in Section 5.2.2.3. The error bars in Figure 5.7 represent changes of 10% in the
intensity threshold used to determine the shrub and gap fractions. A second
method to calculate $F_v$, using hemispherical photographs, gave the same $F_v$ as
the estimation from the aerial photograph. Hemispherical photographs taken on
two other days were then used to calculate the two additional $F_v$ used in Figure
5.7. More details on the determination of $F_v$ can be found in Bewley et al. (2007).
There is a hysteretic relationship between the accumulation and the melt cycles.

![Figure 5.7: Modelled (line) and observed (points with error bars) vegetation fraction as a function of relative snow depth.](image-url)
$F_v/F_{v0}$ decreases relatively fast at the beginning of the accumulation period because this part of the curve is dominated by the bending of the branches (as seen in Figure 5.6b-c). Following this the shape of the curve, which becomes logarithmic, is dominated by the slower burial of the shrub. This logarithmic change in $F_v/F_{v0}$ is also found at the beginning of the melt season, when the shape of the curve is mostly determined by the emergence of shrubs. The steeper part of the melt curve at shallower snow depth is then caused by the fast springing-up of the branches, which the model captures well (rmse = 0.061, $n=3$).

The accumulation curve is a function of snow depth, which slowly buries the base of the shrubs, and of the effective foliage factor (Equation 5.15) which determines how much of the exposed part of the branch can be “seen”. Although $f_{eff}$ can decrease rapidly, the process occurs over a number of timesteps. On the other hand, during melt $S_{rel}$ goes from $1 > S_{rel} > 0$ to 0 in a single timestep for every protruding branch on the assumption that shrubs do not hold snow on the exposed part of their branches when air temperatures are above freezing. In addition, as snow progressively melts, the middle part of some branches is exposed but the base and the tip of the branch are still buried (see the two bent branches on the left in Figure 5.6). As a consequence, there is a hysteresis between the accumulation and melt cycles.

### 5.3.3 Model evaluation against measured albedo

The model is also evaluated against albedo measured at P1 in the WCRB in 2003 (see Chapter 2 for a site description) where shrubs get buried and bent in the winter. Figure 5.8 shows the progressive decline in albedo (measured by radiometers) as shrubs become exposed against an apparent sudden change in snow depth (measured by the ultrasonic depth gauge SR50). What the SR50 actually captures is a shrub springing-up just below the instrument within a 30 minutes timestep.

The modelled albedo, $\alpha$, is calculated as

$$\alpha = (1 - F_v)(1 - F_s)\alpha_g + F_v\alpha_v + F_s(1 - F_v)\alpha_s$$ (5.19)

where $\alpha_g$, $\alpha_v$ and $\alpha_s$ are the respective albedos of the bare ground, the canopy and the snow obtained from field measurements and $F_s$ is the snow fraction, such that

$$F_s = \tanh \left( \frac{SWE}{44} \right)$$ (5.20)
following Essery and Pomeroy (2004a) who found that a hyperbolic tangent fit for $F_s$ closely matched snow surveys in WCRB (Figure 5.9).

Figure 5.10 shows that SBM is able to capture the slow decrease in albedo with increased shrub exposure. The rmse equals 0.07 ($n=27$) but the difference between modelled and measured albedo is relatively constant over the melt season. Two runs with simpler $F_v$ parameterizations were performed in order to demonstrate the necessity to model bending and burying processes. The first one was performed with a modified version of SBM that did not bend the branches but re-calculated $F_v$ taking into consideration the changes in $A_{proj}$ during melt at each timestep. The second version used a fixed $F_v$, set to $F_{v0}$. The fixed $F_v$ parameterization produced the lowest $\alpha$ (rmse = 0.34). As $F_v$ never decreased, the products of both the middle part and the right side of Equation 5.19 always resulted in low albedo values. The simplified “no bending” model provides better results than the fixed $F_v$ version, because the base of the branch is getting buried in the snowpack, but the rmse (0.25) clearly shows the large discrepancy between the original and the modified SBM.

These runs show that a flexible $F_v$ obtained by burying and bending the vegetation considerably improves modelled albedo. Failing to include a flexible vegetation cover parameterization in energy balance models has therefore significant consequences on albedo, and by extension, on the representation of energy exchanges.
Figure 5.8: Changes in albedo and surface height during melt measured over buried tall shrubs.

Figure 5.9: Measured snow cover fraction and average SWE with standard error bars from surveys in the valley in 2003 (open circles) and 2004 (closed circles) and on the plateau in 2004 (crosses) in the Granger basin. The fitted curve is given by Equation 5.20.
Figure 5.10: Changes in albedo with snow depth at P1 in Spring 2003 as measured (circles), modelled with SBM (solid line), with a fixed $F_v$ (dash-dotted line) and with a modified version of SBM that buries but does not bend the shrubs (dash lines).

5.3.4 Sensitivity to parameters

The sensitivity of the modelled $F_v$ and $\alpha$ to the 3 species-specific parameters used in the model (shrub area, Young’s modulus and the mean species snow interception value) was examined and is presented in Figures 5.11 and 5.12. Two runs per parameter were conducted to cover the range 20% above and below the parameter value of the base run. The cumulative effect of changes in multiple parameters was assessed by performing a number of runs where each parameter was randomly generated within the 20% margins used in the single parameter sensitivity. The results of the cumulative tests are not schematically presented because they did not substantially differ from the ones shown in Figures 5.11 and 5.12.

The correlation coefficients (all $> 0.99$) and rmse (all $< 0.01$) between the base run and all the sensitivity runs (including those where the cumulative effect of changes to more than one parameter was tested) show that the model results are robust against parameter variations. These statistics are particularly encouraging.
with regards to the processes occurring in the accumulation cycle as this part of the model has not been evaluated against observations.

Both $A_{\text{ground}}$ and $\overline{S}$ affect the shape of $F_v/F_v0$ in the accumulation cycle (Figure 5.11-a and -c). $\overline{S}$ was calibrated against the relative vegetation fraction value calculated from the aerial photograph (Section 5.3.2) and determined to be 5 kg m$^{-2}$. It is close to but less than the values used in literature for interception of snow by needleleaf trees (between 5.9 and 6.6 kg m$^{-2}$) (Schmidt and Gluns, 1991; Hedstrom and Pomeroy, 1998). Field measurements will be essential to provide empirical data on species of shrubs that intercept snow to better define $\overline{S}$. As $\overline{S}$ affects the accumulation cycle and not the spring-up cycle, it had no effect on the sensitivity of the albedo during snowmelt.

Surprisingly, changes in the Young’s modulus are more significant in the spring-up than in accumulation cycle (Figure 5.11-b). This relationship is shown in Equations 5.4 and 5.10 which state that the amount of bending a branch is subjected to is proportional to its elasticity and height. As a consequence if the elastic modulus is decreased, the branch is less bent in the snowpack and will spring-up sooner for equal snowdepth. The opposite is true if $E$ is increased.

Of the 3 parameters tested, the shrub area, $A_{\text{ground}}$, is the only one that has a significant effect on both the accumulation and the spring-up cycles. The relative vegetation fraction is always sensitive to $A_{\text{ground}}$ except at the end of the melt period when all branches have sprung-up and the shape is only determined by the depth of the snow at the base of the shrub (above $F_v/F_v0 = 0.92$). On the other hand, the most sensitive parts of the curves are the burial and emergence sections (i.e. below $F_v/F_v0 = 0.2$). This is because once the branches have reached their maximum snow holding capacity and $S_{rel} = 1$, $f_{eff}$ is fixed in Equation 5.17 and the change in $PAI$ from then on is solely dependent on the relationship between $A_{\text{proj}}$ and $A_{\text{ground}}$, therefore making the relative vegetation fraction more sensitive to the value of the latter.

Because $A_{\text{ground}}$ affects the exposed vegetation and $E$ affects the shape of the branch, the albedo is sensitive to both in very similar manners. The single parameter sensitivity tests show that decreasing the Young’s modulus or the shrub area lowers the albedo to similar values, closer to observation than the base run.

The cumulative parameters tests showed that the cumulative effect of decreasing $E$ and decreasing $A_{\text{ground}}$ causes albedo values to decrease further.

The sensitivity of SBM to the species-specific parameters, although statistically low, shows that more field work is needed to better define canopy structure parameters.
Figure 5.11: Sensitivity (grey band) of the modelled vegetation fraction (line) to parameter changes of 20% from the base run. Crosses with error bars are measured vegetation fraction values.
Figure 5.12: Sensitivity (grey band) of the modelled albedo (line) to parameter changes of 20% from the base run. Dots are measured albedo values.
5.3.5 Application of SBM in land surface models

Although dynamic vegetation models can simulate shifts in vegetation cover, to the author’s knowledge no model includes an exposed vegetation fraction as calculated in SBM. Yet, vegetation shifts at high latitude require accurate spatial and temporal estimates of vegetation cover if land surface models are to accurately reproduce heat, water vapour and CO$_2$ exchanges between the land and the atmosphere.

As it would be too computationally expensive to couple SBM to a LSM, a parameterization of the vegetation fraction curve obtained with SBM is proposed for use in energy balance models (Figure 5.13) by fitting the equation

$$\frac{F_v}{F_{v0}} = \frac{1}{1 + (S_d/d_h)^m}$$

where $d_h$ is a scaling depth derived from curve fitting for which SBM buries (0.1 m) or reveals (0.25 m) half of the shrub and $m$ is a parameter defining the shape of the curve, given as 3.1 for accumulation and 4.4 for emergence.

This parameterization will be tested and used to inform changes in the vegetation fraction in JULES in the next chapter and in the distributed 3-source model introduced in Chapter 7.

5.4 Summary

Shrubs get buried and / or bent during the winter to emerge rapidly during the snowmelt season. This behaviour has significant consequences on the energy balance of shrub-tundra landscapes not only because of the difference in albedo between snow and vegetation covers but also with regards to heat and water exchanges with the atmosphere.

In this chapter, the physics of a simple shrub bending model (SBM) were presented. The model was then evaluated against measured vegetation fractions and observed changes in albedo during snowmelt. The results warranted confidence in parametrising the modelled vegetation fraction for use in energy balance model.

In the following chapter, the $F_v$ parameterization is implemented in JULES in order to improve the representation of albedo over shrub-tundra landscapes in the model.
Figure 5.13: Modelled (solid line), observed (points with error bars) and parametrised (dashed line) vegetation fraction.
Chapter 6

A new snow cover fraction to assess the effect of shrubs on tundra hydrometeorology
6.1 Introduction

Results from Chapter 4 led to the conclusion that a better representation of shrubs processes was needed in JULES if the effect of shrub-tundra expansion was to be investigated with the model. In the previous chapter, a high resolution model, able to replicate the bending of shrubs under the weight of snow, was proposed and used to develop a parameterization of the exposed vegetation fraction for use in large scale models. Here, this parameterization is implemented in all the versions of JULES tested in Chapter 4 and evaluated against streamflow. The version of the model giving the best overall performance is then used to investigate the potential effect of shrub expansion on hydrometeorology in the Abisko basin.

6.2 A new snow cover fraction parameterization for the shrub tile

The albedo parameterization in JULES (Equation 3.13) assumes a complementary relationship between the snow covered and the snow-free fraction. In addition, in the snow cover parameterization (Equation 3.15), the relationship between the snow depth and the canopy height, represented by $10z_0$, suggests that snow buries vegetation but does not bend it. The effect of these assumptions were shown to lead to albedo being underestimated by up to 40% over a melting heterogeneous surface (Figure 5.10). In addition, the snow-free fraction in the model does not distinguish between exposed vegetation and bare ground and the snow fraction does not distinguish between snow not masked by protruding vegetation and snow in unvegetated areas.

In the previous chapter a model that reproduces the bending of branches under the weight of snow was introduced. The exposed vegetation fraction was parametrised in Equation 5.21, in order to describe the non-linear relationship between snow depth and exposed vegetation. By implementing $F_v$ in JULES, a new parameterization for the fraction of the tile with snow not masked by vegetation is proposed

$$F_{sv} = F_s(1 - F_v)$$

(6.1)

where $F_v$ and $F_s$ are given by Equations 5.21 and 5.20 respectively. Assuming that the albedo of shrubs equals that of bare ground, the albedo for the shrub tile can be parametrised by re-interpreting Equation 3.13 as
\[
\alpha_{\text{tile}} = F_{sv}\alpha_s + (1 - F_{sv})\alpha_{st} \tag{6.2}
\]

The albedo for non-shrub tiles remains as described in Section 3.4.

### 6.3 Evaluation of the new shrub tile albedo

Following Chapter 4 where JULES was shown to perform poorly over the shrub site, the new snow cover fraction parameterization is evaluated against the same data. The new parameterization was implemented in every version of the model evaluated in Chapter 4, for which short descriptions are presented in Table 6.1 for convenience. The original snow cover fraction (Equation 3.15) is hereafter referred to as the old SCF while Equations 6.1 and 6.2 are referred to as the new SCF.

<table>
<thead>
<tr>
<th>JULES scheme</th>
<th>Scheme characteristics</th>
</tr>
</thead>
<tbody>
<tr>
<td>JULES-Ctl</td>
<td>Original JULES runoff scheme.</td>
</tr>
<tr>
<td>JULES-SD</td>
<td>Water in excess of saturation is passed to the layer below rather than the layer above as in JULES-Ctl.</td>
</tr>
<tr>
<td>JULES-TOP</td>
<td>Runoff is calculated by assuming a relationship between runoff rate, topography and water table depth.</td>
</tr>
<tr>
<td>JULES-TOP-NY</td>
<td>Combination of JULES-NY and JULES-TOP.</td>
</tr>
</tbody>
</table>

Table 6.1: Description of the 5 hydrology schemes used in Chapter 4.

#### 6.3.1 Effect of the new snow cover fraction in Abisko

<table>
<thead>
<tr>
<th>JULES scheme</th>
<th>Old SCF</th>
<th>New SCF</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>rmse</td>
<td>bias</td>
</tr>
<tr>
<td>JULES-Ctl</td>
<td>21.3</td>
<td>-1.24</td>
</tr>
<tr>
<td>JULES-SD</td>
<td>17.4</td>
<td>-1.35</td>
</tr>
<tr>
<td>JULES-NY</td>
<td>16.9</td>
<td>-1.26</td>
</tr>
<tr>
<td>JULES-TOP</td>
<td>23.3</td>
<td>-1.68</td>
</tr>
<tr>
<td>JULES-TOP-NY</td>
<td>22.7</td>
<td>-2.11</td>
</tr>
</tbody>
</table>

Table 6.2: Statistics of observed streamflow against modelled runoff with the new SCF in the Abisko catchment. Observed streamflow against modelled runoff with the old SCF are reproduced from Table 4.2 for convenience.
Figure 6.1: 10-year catchment average observed streamflow (dots) against modelled runoff with the old (red line) and new (black line) SCFs in the Abisko catchment.
The performance of each model with each SCF against observations is shown in Figure 6.1. Table 6.2 shows that 11 out of 15 statistics are improved in the Abisko catchment with the implementation of the new SCF. Figure 6.2 shows the new and old SCFs and their associated shrub tile albedo over a single gridbox in the Abisko catchment. For simplicity, output from a single gridbox is presented. The chosen gridbox is the most representative of shrub processes as it is the one with the largest shrub fraction in the catchment. The new SCF is up to 80% higher than the old SCF (Figure 6.2-a) during the accumulation cycle and up to 60% during melt, which leads to the tile albedo being up to 0.49 higher with the new scheme (Figure 6.2-b). This increase in albedo generates slower melt rates at the onset of the snowmelt season, which considerably improves snowmelt rates in the three versions of the model that are dominated by surface runoff (JULES-Ctl-$F_{sv}$, JULES-TOP-$F_{sv}$ and JULES-TOP-NY-$F_{sv}$), especially at the beginning of the snowmelt season. Although JULES-CTL-$F_{sv}$ suffers a sharp decrease in runoff at the end of June / beginning of July not seen in observations, slower melt rates at the beginning of spring allow more water to infiltrate the soil column before saturation is reached. As a consequence, the soil is less dry in the summer and more rainfall-induced subsurface runoff is generated. In addition, as the new SCF allows snow to bury the vegetation and to stay on the ground for longer, less water is lost to soil and canopy evaporation in spring and summer, which decreases on average, between all versions, by 27% over the basin. This is corroborated by the reduction in bias for all versions which indicates that the models are losing less moisture with the new SCF.

Figure 6.3 shows the difference in snow depth, net radiation and turbulent fluxes in the basin between JULES-Ctl and JULES-Ctl-$F_s$. Although there seems to be little difference in snow depth from March to July there are some significant differences in the net radiation and turbulent fluxes. These differences come from the relationship between snow depth and canopy height in the respective SCF. With the new SCF, $F_{sv} = 1$ when snow depth equals or is greater than canopy height, $h_c$. On the other hand with the calculation of the snow cover fraction in Equation 3.15, $F_s$ only equals 0.5 when $S_d = h_c$. While this difference does not matter between October and March when there is little or no incoming shortwave radiation, this change in the albedo leads to an average reduction in net shortwave radiation of 31 W m$^{-2}$. This in turn causes less energy to be available for melt which is reflected in the hydrograph where flow rates are smaller from the beginning of May to the end of July.
Figure 6.2: Comparison, between JULES-Ctl (red) and JULES-Ctl-$F_s$ (black) from 1989 to 1999, of the relationship between snow depth and (a) snow cover fraction (b) albedo, over the shrub tile (shrub height = 1.5 m) at one point in the Abisko catchment.
Figure 6.3: 10 years catchment average of the snow depth, net radiation, latent heat fluxes and sensible heat fluxes with JULES-Ctl (red line) and JULES-Ctl-$F_s$ (black line).
Both latent and sensible heat fluxes are smaller with the new SCF. The change in $H$ is particularly significant with regards to land surface - atmosphere interactions because $H$ is now negative from April to mid-June. This means that, by allowing shrubs to get buried in the snowpack, introducing $F_v$ in albedo calculation causes the catchment to move from heat source to heat sink in April (Figure 6.4) and decreases average turbulent fluxes from the land surface to the atmosphere between 1 April and 1 July by 26 W m$^{-2}$. This shows that omitting the bending and burying of the vegetation in land surface models could cause modelled predictions to overestimate the positive feedback between land surface changes, such as shrub expansion, and climate warming.

Going back to the intercomparison of the 10 versions of JULES, the two models that performed best with the old SCF are the only two models whose efficiency and rmse have worsened with the implementation of the new SCF. This occurs because both JULES-SD and JULES-NY allow more water in the soil column. As a consequence snowmelt is transformed predominantly into subsurface flow which delays runoff compared to the control run. As is seen Figure 6.2, one of the effects of the new SCF is also to increase the shrub tile albedo and therefore to delay runoff. The combined effects of delaying both snowmelt and runoff in JULES-SD-$F_s$ and JULES-NY-$F_s$ leads to peak flow being too late.
6.3.2 Effect of the new SCF in the Torne and Kalix basins

<table>
<thead>
<tr>
<th></th>
<th>Old SCF</th>
<th></th>
<th>New SCF</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>rmse</td>
<td>bias</td>
<td>efficiency</td>
<td>rmse</td>
</tr>
<tr>
<td>JULES-Ctl</td>
<td>640</td>
<td>62.04</td>
<td>-0.30</td>
<td>608</td>
</tr>
<tr>
<td>JULES-SD</td>
<td>456</td>
<td>46.93</td>
<td>0.34</td>
<td>461</td>
</tr>
<tr>
<td>JULES-NY</td>
<td>390</td>
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<td>0.52</td>
<td>388</td>
</tr>
<tr>
<td>JULES-TOP</td>
<td>659</td>
<td>45.11</td>
<td>-0.38</td>
<td>626</td>
</tr>
<tr>
<td>JULES-TOP-NY</td>
<td>364</td>
<td>17.71</td>
<td>0.58</td>
<td>363</td>
</tr>
</tbody>
</table>

Table 6.3: Statistics of modelled against observed streamflow with the new SCF in the Torne and Kalix basin. The statistics with with the old SCF were reproduced from Table 4.1 for convenience.

The performance of each model with each SCF against observations for TKB is shown in Figure 6.5 while results from statistical analyses are presented Table 6.3.

Unlike in Abisko, only 6 out of 15 statistics are improved and two remain unchanged. The new SCF has no or very little effect on the basin south of a diagonal going from $\sim 67^\circ 5'N \ 19^\circ 25'E$ to $\sim 68^\circ 75'N \ 22^\circ E$ where shrub cover is limited. North of this diagonal, the effect of the new SCF on streamflow is the same as on runoff in Abisko therefore what was mentioned in Section 6.3.1 is not repeated here.

The main effect on the streamflow in TK is that melt from the shrubland in the north of the basin is delayed and starts reaching the mouth of the river up to 10 days later than it did with the old SCF. For the surface runoff-dominated schemes (JULES-Ctl-\(F_s\), JULES-TOP-\(F_s\) and JULES-TOP-NY-\(F_s\)), the delay in snowmelt with the new SCF on the shrub tiles causes flow rates to slow down between melt from the boreal forest in the south of the basin and melt from the shrubs in the north. However, as was seen in the previous section in Abisko, because melt rates are now slower the melt season is prolonged, the soil is wetter in Summer and the model is able to capture the rainfall-induced flow season. As a consequence performance loss in the middle of the snowmelt season is compensated for by a better simulation of rainfall-runoff such that the statistical results hardly change between the two SCF.
Figure 6.5: 10-year average observed (dots) and modelled streamflow at the mouths of the Torne and the Kalix rivers with the old (red line) and new (black line) SCFs.
6.3.3 Comparison of the performance of the new SCF in Abisko and TK

<table>
<thead>
<tr>
<th>Model</th>
<th>Abisko</th>
<th>TK</th>
</tr>
</thead>
<tbody>
<tr>
<td>JULES-Ctl</td>
<td>7</td>
<td>9</td>
</tr>
<tr>
<td>JULES-SD</td>
<td>3</td>
<td>5</td>
</tr>
<tr>
<td>JULES-NY</td>
<td>2</td>
<td>= 3</td>
</tr>
<tr>
<td>JULES-TOP</td>
<td>10</td>
<td>10</td>
</tr>
<tr>
<td>JULES-TOP-NY</td>
<td>9</td>
<td>= 1</td>
</tr>
<tr>
<td>JULES-Ctl-$F_{sv}$</td>
<td>1</td>
<td>7</td>
</tr>
<tr>
<td>JULES-SD-$F_{sv}$</td>
<td>5</td>
<td>6</td>
</tr>
<tr>
<td>JULES-NY-$F_{sv}$</td>
<td>4</td>
<td>= 3</td>
</tr>
<tr>
<td>JULES-TOP-$F_{sv}$</td>
<td>6</td>
<td>8</td>
</tr>
<tr>
<td>JULES-TOP-NY-$F_{sv}$</td>
<td>8</td>
<td>= 1</td>
</tr>
</tbody>
</table>

Table 6.4: Order of the performance of the models from 1 (best) to 10 (worst) based on efficiency alone for the Abisko basin and the Torne and Kalix rivers

The 10 versions of the model are ranked according to efficiency alone in Table 6.4 for the Abisko and the Torne and Kalix basins. While the ranking shows that the weakest version in both basins is JULES-TOP, there is no clear “best” performing model. The two versions using JULES-TOP-NY perform well in TK but are at the bottom of the table in Abisko to within one version. Equally JULES-Ctl-$F_{sv}$, which is at the top of the table in Abisko, is number 7 in TK with a negative efficiency coefficient. Potential sources of errors in the models that were not addressed in the implementation of the new SCF are discussed in Section 4.9.

JULES-NY would be the “best” performing model if efficiency was taken as the sole criterion for model performance, being the second and third model in Abisko and TK respectively. However, the statistics mask large inefficiencies in the model, notably the lateness of JULES-NY with regards to the onset of snowmelt-induced discharge and the failure to time peak flow correctly in TK. Yet, JULES-NY has the advantage of containing only one modification to the original model, thus minimising the number of new parameters. The development of ever more complex models has led many scientists to express the need for simplicity in model structures on the assumption that although more complex models may lead to improved performances, these improvements may also be the results of compensating for errors in other parts of the model. (e.g. see Nash and Sutcliffe, 1970; Beven, 1993; Perrin et al., 2001).

In addition, equifinality occurs in this study where, for example, adding the NY
scheme causes a delay in runoff production while adding the new SCF causes a slowing down of snowmelt rates. As a consequence, both schemes lead to a delay in snowmelt runoff. While the new parameterizations perform well individually over the Abisko basin, combining both schemes reduces model performance and causes JULES-NY-$F_{sv}$ to miss the onset of the spring runoff by approximately three weeks. With only streamflow rates being available for evaluation, it is unclear which parameterization addresses actual model deficiency and / or compensates for errors in the other parameterizations. Besides, although some sensitivity tests were performed (see Chapter 4), all the added parameters were used with their default values, i.e. they were not calibrated for use in the Abisko and TK basins, which can lead to more uncertainties.

JULES-Ctl-$F_{sv}$ is the best performing model in Abisko. Like JULES-NY, it only contains a single modification to the original model but, unlike JULES-NY, the modification included in JULES-Ctl-$F_{sv}$ was developed specifically to address deficiencies in the treatment of shrub-tundra processes in JULES. Furthermore the new SCF was evaluated against streamflow in Abisko and the TKB but also against the vegetation fraction and the albedo at a third site in the Wolf Creek Research Basin. Because this study focuses on the influence of shrubs on snowmelt energetics, a model set-up that performs well over Abisko is essential in order to investigate the impact of shrub expansion on the hydrometeorology of shrub-tundra landscapes. As a consequence JULES-Ctl-$F_{sv}$ will be used in the following section to perform further investigations into shrub-specific processes in the catchment.

6.4 Assessing the impact of shrub cover in Abisko

Following the land classification given in the PILPS2e experiments (see Figure 2.14), approximately 63% of the land cover in the Abisko catchment can be described as open or closed shrubland, 35% as unvegetated land including 11% lakes and < 2% grassland or wooded grassland. As shrub-tundra is expected to encroach on tundra landscapes, conducting an investigation by removing shrubs from the landscape would allow the assessment of shrub-specific effects on hydrometeorology. Such studies can help reduce the uncertainties with regards to what may happen to tundra threatened by shrub encroachment at high latitudes. In order to perform the investigation, the JULES-Ctl-$F_{sv}$ runs are compared against the no-shrub runs. The following minimum necessary number of modifications to the input file were implemented in order to minimise errors due to
model set-up rather than to differences between model outputs:

- The shrub cover was replaced by 0.1 m tall grass. Because Equations 5.20, 6.1 and 6.2 were specifically implemented for the shrub tile, running the model with grass instead of shrubs means that the tile albedo is calculated with Equations 3.13 and 3.15.
- For each plant functional type JULES requires that the canopy model described in Section 3.7.3 be switched on or off. The main purposes of the canopy model are to (a) allow snow interception by vegetation (b) limit sublimation from the snow on the ground (c) promote sensible heat exchange between the canopy and the snow on the ground. If the canopy model is off, snow is assumed to cover the surface and turbulent fluxes are calculated with Equations 3.5 and 3.6 assuming $T_s = \text{snow temperature}$. This set-up was deemed more appropriate for grass which is short and quickly gets buried in the snowpack and, as a consequence, the canopy model was switched off.

### 6.4.1 Results

Figure 6.6 shows modelled runoff with the current vegetation cover against the no-shrub run. In agreement with the snow depth curves (Figure 6.7) which show that snowpack ablation occurs earlier in the grass run, results suggest that shrubs in the Abisko catchment delay runoff by approximately two weeks. Although there is little difference in snowdepth between the shrub run and the grass run, it is important to note that the snowpack is deeper in the grass run which contradicts past studies that showed that snow trapping by shrubs leads to deeper snow than in open tundra environments (Sturm et al., 2001a, 2005a; Pomeroy et al., 2006; Bewley et al., 2010).
Early snowmelt in the grass run leads to larger net radiation in May and June. Whereas in May much of the additional radiation is used for snowmelt, in June most of it contributes to larger latent and sensible heat fluxes from the surface to the atmosphere. Like in the previous paragraph, these results contradict the studies referenced above where sensible heat fluxes from shrub-tundra were found to be greater than those from open environments.

In the next two sections, the disagreement between the results in Figure 6.7 and those reported by previous researchers will be discussed in order to determine if the model results are accurate reproductions of shrub-tundra processes or artefacts of model parameterizations.

### 6.4.1.1 Issues with the canopy model

The canopy model was initially developed for use in coniferous forests where the combination of the large exposed surface area of the snow grains and the roughness length of the trees leads to high sublimation rates for intercepted snow (Essery et al., 2003b; Essery and Clark, 2003). On the other hand, sublimation for the snow on the ground is suppressed and melt is delayed by the reduced transmission of solar radiation through the canopy and by wind shelter. The
Figure 6.7: 10-year average modelled snowdepth, net radiation, latent and sensible heat fluxes under the current vegetation cover (black) and without shrubs (blue).
model was then made available for all plant functional types but no adaptation for type-specific processes (notably shrubs) was introduced. In Chapter 5, the author hypothesised that shrubs also intercept snow but, instead of keeping the snow on their branches like conifers keep snow on their needles, snow interception in shrubs causes them to bend under the weight of the snow and to get buried under the snowpack in winter. As a consequence, where shrubs are buried under the snowpack, sublimation is expected to occur at rates similar to those at open-tundra sites where up to 40% of the snow water equivalent can be lost to sublimation (Pomeroy and Gray, 1995). Moisture fluxes from the canopy are calculated according to Equation 3.40 reproduced below for convenience

$$E = \frac{\rho}{r_a + r_i}[Q_{\text{sat}}(T_c, P) - Q_a].$$

(6.3)

When the canopy is completely snow covered, $r_i$ is set to 0. $r_a$ partly depends on surface roughness therefore the taller the vegetation, the more moisture fluxes from the canopy are diagnosed. While this formulation is appropriate for coniferous forests where a large fraction of the intercepted snow is lost to sublimation (Pomeroy and Gray, 1995), the roughness length of a surface where shrubs are buried under the snowpack is in effect the same as that of a snowpack burying grass. As the model is intercepting snow rather than intercepting snow and burying shrubs, more sublimation is produced from the shrub tile than from the grass tile even if $S_d > h_c$ ($h_c = 0.1$ m on the grass tile and 1.5 m on the shrub tile).

Although the effect of buried shrubs on albedo was implemented in Equation 6.1, parametrising the effect of bent rather than protruding shrubs on sublimation in JULES would require major changes to the structure of the model that are beyond the scope of this study.

Another issue with the canopy model is that all the parameters used in the model have been derived from extensive field studies on needleleaf trees (Schmidt and Gluns, 1991; Hedstrom and Pomeroy, 1998). As was seen in Chapter 5, interception of snow by shrubs can be likened to interception by needleleaf trees only to a degree and transferring one to the other does not simply require parameter adjustments in an existing model but necessitates a new model altogether.

Another approach would be to remove the interception from the canopy model on shrub tiles. Figure 6.8 shows that this method does lead to slightly deeper snow in the shrub run than in the grass run because the sublimation from the shrub tile is, in effect, removed. This method would be appropriate if sublimation from snow on the ground was allowed when $S_d > h_c$. Instead, moisture fluxes are calculated from the canopy with Equation 6.3. As a consequence, if snow
interception is removed, sublimation equals 0 even if $S_d > h_c$ and regardless of whether the canopy is fully exposed or buried within the snowpack. Therefore, removing interception does not offer a better solution for modelling shrub-tundra processes than keeping it. Essentially, the problem with transferring the canopy model for use in shrub-tundra environments is that the canopy model was developed to deal with vertical processes (snow interception and throughfall in boreal forests) whereas shrub-tundra is determined by both vertical (interception of snow, bending and burying) and horizontal processes (trapping of wind-blown snow from open areas, heat advection between shrubs and adjacent open areas). Whether JULES is used with or without a canopy model and with or without interception, the structure of the model does not allow it to simulate the transition between the processes occurring in erect shrub environments and those occurring in bent shrub environments.

### 6.4.1.2 Limitations of the snow cover fraction parameterization

The second significant difference between the current vegetation run and the no-shrub run is the snow depth, which is reflected in the timing of the runoff. Yet, all else being equal, one would expect the snow cover fraction of the grass run to be higher than that of the shrub run as the snowpack is deeper in the former than in the latter. Instead, the lower SCF of the grass tile (Figure 6.9) generates earlier snowmelt and, as a consequence, earlier runoff. The new SCF was parametrised specifically for use in shrub-tundra environments. On the other hand, the old SCF is a simple function used for all other tiles. As can be seen in Figure 6.9, using the shrub-specific SCF alongside the SCF for all other tiles clearly leads to inconsistencies between shrub and non-shrub tiles whereupon the albedo of the shrub tile is higher than that of the grass tile, even...
for snow depths where all vegetation types are buried under the snowpack. The solution proposed by Roesch et al. (2001) is to implement specific SCFs for each functional type. While this work is beyond the scope of this study, further work into landscape- and landcover-specific snow cover fraction parameterizations may identify more tile-specific SCFs that would allow JULES to investigate snow-vegetation interactions at high latitudes.

6.5 Discussion and conclusion

In Chapter 4, the performance of JULES against measured streamflow in an Arctic basin was improved by parametrising the sub-grid soil moisture heterogeneity. In this chapter, runoff simulations in a shrub-tundra sub-basin were improved by addressing sub-grid snow cover heterogeneity. However, further investigations have identified issues with the canopy model and its treatment of moisture fluxes and inconsistencies between the snow cover fraction parameterizations. Essentially, both the canopy model and the new SCF aim to represent three coexisting surfaces: the vegetation, the snow and the bare ground. In boreal forests, for
which the canopy model was implemented, the interaction between those surfaces is vertical. In shrub-tundra environments, the interaction is horizontal. To simulate sub-tile horizontal heterogeneity using the current structure of the model would require the addition of many parameters. The solution proposed in many LSMs (e.g. SiB, Seller et al., 1986; CLASS, Verseghy et al., 1993; ISBA, Douville et al., 1995) is to calculate separate energy balances for vegetation and for bare ground or snow. Blyth et al. (1999) implemented such a scheme in MOSES for use over a fallow savannah site with tall shrubs and grass and found that the model performed significantly better than the version calculating a single energy balance. However, making such changes in JULES would require a major reorganisation of the model structure.

The results presented in this chapter show that investigations into the potential effect of shrub expansion on the hydrometeorology of a shrub-tundra basin can not be performed with JULES in its current structure. As a consequence, in order to address the aim and objectives of this study stated in 1.8 and to quantify the need for a more sophisticated model, a structurally simpler model but with more sophisticated physics adapted for shrub-tundra environments needs to be developed. This model is proposed in the next chapter.
Chapter 7

Distributed 3-Source Model (D3SM)
7.1 Introduction

JULES represents sub-grid heterogeneity through a tiling scheme. The energy balance for each tile is calculated separately and areally weighted to produce the gridbox energy balance. As a consequence, the model structure does not allow for the advection of heat to the snowpack from shrub canopies or bare ground patches even though discontinuous snow cover is known to enhance snowmelt \((e.g.\) Liston (1995); Neumann and Marsh (1998); Pomeroy et al. (2003)).

In order to provide a better representation of sparse vegetation processes, Blyth et al. (1999) proposed a dual-source scheme, for application in the savannah, in which separate energy balances for the vegetation and the bare ground tiles are solved simultaneously. This model was later adapted by Bewley et al. (2010) for application in shrub-tundra environments. This chapter presents further developments to the Blyth et al. (1999) and Bewley et al. (2010) model by introducing a third source and by applying the model over a distributed landscape. The model will then be evaluated against manual snow depth measurements and measured sensible and latent heat fluxes from the Granger basin (see Section 2.2.1 for site descriptions).

7.2 The surface radiation and energy balance

The physics of the 3-Source Model (3SM) are as described in Section 3.3, but adapted to calculate separate radiation and energy balances for the snow-free ground, snow and vegetation surfaces, hereafter designated by the subscripts \(g\), \(s\) and \(v\) respectively (see Figure 7.1). In order to calculate separate energy balances, Equations 3.1, 3.2, 3.5 and 3.6 need to be partitioned into the three sources such that

\[
R_v = H_v + LE_v
\]  
(7.1)

\[
R_g = H_g + LE_g
\]  
(7.2)

\[
R_s = H_s + LE_s + M
\]  
(7.3)

The ground heat flux in Equation 3.1 is assumed to be negligible and, in the absence of a soil model, was removed from Equations 7.1 to 7.3. The radiation and the turbulent fluxes are calculated per unit area of the sources unless stated otherwise.
7.2.1 Calculation of the radiative fluxes

In order to calculate the radiative fluxes, the elements on the right side of Equation 3.2 need to be partitioned into

\[ SW_s = (1 - F_v + \tau F_v)(1 - \alpha_s)SW_\perp \]  \hspace{1cm} (7.4)

for the snow tile,

\[ SW_v = (1 - \alpha_v)SW_\perp \]  \hspace{1cm} (7.5)

for the vegetation tile and

\[ SW_g = (1 - F_v + \tau F_v)(1 - \alpha_g)SW_\perp \]  \hspace{1cm} (7.6)

for the ground tile, where \( \tau \) is the canopy transmissivity, \( F_v \) is the vegetation fraction given by the parameterization in Equation 5.21 and \( \alpha_v, \alpha_g \) and \( \alpha_s \) are vegetation, bare ground and snow albedos determined from measurements to be 0.11, 0.15 and 0.85 respectively (Bewley, 2006). Multiple reflections between the vegetation and the ground or snow are neglected.

On the vegetation tile, the longwave radiation is emitted both from the top of the canopy to the atmosphere and from under the canopy to the ground or snow such that
\[ \text{LW}_v = \text{LW}_\downarrow - 2 \sigma T_v^4 + \sigma T_s^4 + \sigma T_g^4 \]  

(7.7)

where \( T_v \) is the vegetation temperature. Absorption of longwave radiation from the adjacent vegetation is accounted for in the calculation of the net longwave radiation for the snow and vegetation surfaces:

\[ \text{LW}_s = (1 - F_v) \text{LW}_\downarrow + F_v \sigma T_v^4 - \sigma T_s^4 \]  

(7.8)

\[ \text{LW}_g = (1 - F_v) \text{LW}_\downarrow + F_v \sigma T_v^4 - \sigma T_g^4 \]  

(7.9)

where \( T_s \) is the snow temperature and \( T_g \) is the ground temperature.

### 7.2.2 Calculation of the turbulent fluxes

Like the radiation balance components, the turbulent fluxes are calculated separately for each source. The sensible heat fluxes at the snow, bare ground and vegetation surface are respectively

\[ H_s = \frac{\rho C_p}{r_{as}} (T_s - T_c) \]  

(7.10)

\[ H_g = \frac{\rho C_p}{r_{ag}} (T_g - T_c) \]  

(7.11)

\[ H_v = \frac{\rho C_p}{r_{av}} (T_v - T_c) \]  

(7.12)

where \( T_c \) is the temperature of the canopy air space and \( r_{as}, r_{ag} \) and \( r_{av} \) are the aerodynamic resistances for snow, ground and vegetation to be defined in section 7.2.3.

For conservation of energy, the sensible heat flux to the atmosphere is

\[ H = (1 - F_s) H_g + F_s H_s + F_v H_v \]  

(7.13)

where \( F_s \) is the snow cover fraction given by Equation 5.20.

Moisture fluxes over the snow \( (E_s) \) and ground \( (E_g) \) surfaces are calculated as

\[ E_s = \frac{\rho}{r_{as}} [Q_{sat}(T_s, P_s) - Q_c] \]  

(7.14)

and

\[ E_g = \psi \frac{\rho}{r_{ag}} [Q_{sat}(T_g, P_g) - Q_c] \]  

(7.15)
where $Q_c$ is the canopy air space specific humidity. Transpiration from the dormant vegetation is assumed to be negligible (i.e $LE_v = 0$).

For conservation of mass, the total moisture flux going to the atmosphere is

$$E = F_s E_s + (1 - F_s) E_g$$ (7.16)

The five unknowns ($T_c$, $T_g$, $T_s$, $T_v$ and $Q_c$) are found by solving the five Equations 7.1 to 7.3, 7.13 and 7.16 simultaneously, using LU (Lower and Upper triangular matrices) decomposition following Press et al. (1996). The model first solves the equations assuming no snowmelt. If the model diagnoses the snow surface temperature to be greater than 0°C, the solution is repeated by assuming $T_s = 0°C$ and the residual of the energy balance is used to melt the snow at the rate of $L_f S_M$.

### 7.2.3 Aerodynamic resistances and roughness lengths

Aerodynamic resistances, essential to the calculation of the sensible and latent heat fluxes, are bulk boundary layer descriptors that describe the efficiency of turbulent transfer from the surface to the atmosphere. Turbulent transport beneath a canopy is poorly understood and as a consequence many parameterization of aerodynamic resistances encompassing various degrees of complexity exist (e.g. Dolman (1993); Huntingford et al. (1995); Blyth et al. (1999); Zeng et al. (2004)). The formulations in 3SM are adapted from Blyth et al. (1999).

The aerodynamic resistance for energy exchange between the canopy air space and reference height for temperature, $z_t$, is given by

$$r_{aa} = \frac{1}{k u_*} \ln \left( \frac{z_t}{d} \right)$$ (7.17)

where $k$ is the von Karman constant, $u_*$ is the friction velocity and $d$ is the displacement height, taken to be 2/3 of the canopy height. The aerodynamic resistance for exchanges between the vegetation and the canopy air space is given by

$$r_{av} = \frac{1}{k u_*} \ln \left( \frac{d}{z_{0v}} \right)$$ (7.18)

where $z_{0v}$ is the roughness length of the vegetation, assumed to be 1/10 of the canopy height.

Bewley et al. (2010) found that for understorey resistances, the form used in Equations 7.17 and 7.18 leads to an increase in sublimation with increasing veg-
etation cover rather than suppressing turbulent transport from the underlying snow. They proposed a modified form, which adapted for a 3-source model gives:

\[
\frac{1}{r_{as}} = \left[ \frac{1 - F_v}{\ln(d/z_{0s})} + F_s C \right] k u_s
\]

(7.19)

\[
\frac{1}{r_{ag}} = \left[ \frac{1 - F_v}{\ln(d/z_{0g})} + F_s C \right] k u_s
\]

(7.20)

where \( C \) is a dense canopy exchange coefficient given the value of 0.004 (Zeng et al., 2002) and \( u_s \) is the friction velocity such that

\[
u_s = \frac{k u}{\ln(z_u/z_0)}
\]

(7.21)

where \( u \) is wind speed, \( z_u \) is the reference height for wind speed and \( z_0 \) is the momentum roughness length for which, as for the aerodynamic resistances, there exists a number of different formulations. For \( F_v = 0 \) or 1, Equations 7.19 and 7.20 are reduced to the usual logarithmic forms for resistances over homogeneous surfaces. Following Mason (1988), an “effective” roughness length, intended to represent a spatial average in a heterogeneous terrain, is calculated by weighting the individual roughness values of the 3 sources such that

\[
z_0 = z_u \exp \left\{ - \left[ \frac{F_v}{\ln^2(z_u/z_{0v})} + \frac{F_s(1-F_v)}{\ln^2(z_u/z_{0s})} + \frac{(1-F_v)(1-F_s)}{\ln^2(z_u/z_{0g})} \right]^{-1/2} \right\}
\]

(7.22)

### 7.3 Model evaluation

#### 7.3.1 Initial conditions

Snow depth and turbulent fluxes measured in Spring 2003 and Spring 2004 were used to evaluate 3SM. Initial snow depth and vegetation heights were prescribed from manual measurements. The snow-free vegetation fraction at GB2 was calculated from an aerial photograph as described in Section 5.3.2 and \( \tau \) was set at 0.57 following Bewley et al. (2010) who estimated canopy transmissivity from hemispherical photography. As \( F_{v0} \) and \( \tau \) were not estimated at P2, they were attributed the values of 0.1 and 0.9 respectively, which were chosen to reflect the fact that vegetation at the site is short and sparse. In the absence of a soil model and of any actual soil moisture measurements, values for the soil moisture factor \( (\psi \text{ in Equation 7.15}) \) at P2 and GB2 were calibrated against the heat fluxes after snowmelt and were found to be 0.1 and 0.2 respectively.
7.3.2 Results

Model performance at a point is evaluated against manual snow measurements and automatic snow and turbulent flux measurements at GB2 and P2 (Figures 7.2 to 7.5). The contribution of the individual sources to the transfer of energy is shown in Figures 7.6 and 7.7 for 2003 only as, although the magnitude of the fluxes are different in 2004, the process behind the distribution of the fluxes between the sources is the same each year.

Snow depth simulations (panel b in Figures 7.2 to 7.5) show that the model is able to reproduce snowmelt patterns over a heterogeneous landscape. Following an initial period when the model melts snow slightly faster than observed, 3SM not only closely reproduces the date at which the ground becomes snow-free but also the rate at which snowmelt occurs. The model is also able to reproduce the direction and scale of energy exchanges between the surface and the atmosphere over a dynamic landscape (panels c and d in Figures 7.2 to 7.5). For example, the large positive sensible heat fluxes at GB2 start while there is still snow on the ground but as soon as $F_v$ increases (Figure 7.2a), showing that the model can “see” protruding vegetation.

Diurnal cycles are also reflected in the model simulations. At P2, $H_g$ is positive during daytime from the beginning of the run (Figure 7.7c) but $H_v$ (Figure 7.7e) is not, meaning that the model recognises a small snow-free fraction but no protruding vegetation or, in other terms, $F_s \neq 1$ but $F_v = 0$ (Figure 7.3a). At GB2, $H$ (Figure 7.2d) is positive during daytime owing to positive contributions from $H_v$ (Figure 7.6e) and negative at night owing to negative contributions from both $H_s$ and $H_v$ (Figure 7.6d-e).

Heat advection from the shrub tile dominates snowmelt at GB2 where melt rates are on average 64% larger than at P2 in 2003 and 56% larger in 2004. This is in agreement with Pomeroy et al. (2006) who reported that melt rates at a tall shrub site were generally higher than those over a sparse canopy site because warm shrub branches promote melting by advecting heat to the surrounding snow. Thus, although the initial SWE is approximately twice as large at GB2 as at P2 (187 and 88 mm respectively in 2003 and 140 and 68 mm in 2004), 3SM diagnoses complete melt to occur at GB2 only five days after complete melt at P2.

On the other hand $H$ at P2 (Figures 7.3d and 7.5d) is predominantly negative until the snowpack has mostly disappeared, reflecting the fact that the vegetation at P2 is sparser and shorter than at GB2. The increase in the magnitude of $H_v$ (Figure 7.6e) is due to the varying $F_v$ which increases as the snow pack becomes
shallower. While the fluxes are close to measurements when $F_s > 0$, the model tends to overestimate them when the snow is very shallow or completely melted. Adding a soil model to inform 3SM of the soil moisture content may resolve this issue.

Bewley et al. (2010) performed similar runs with a dual-source (snow and vegetation) model and found that, in the absence of a separate ground tile, their model melted the snowpack too early at GB2 because it could not reproduce late-lying snow patches. Also, from days 113 to 116, sensible heat fluxes at P2 were of opposite sign from the observations, but of similar magnitude, meaning that the difference between observed and modelled $H$ on day 115 was approximately 250 W m$^{-2}$. Results presented in this section showed that including a third source (ground) produces a more accurate representation of the timing of snowmelt over a dynamic landscape.
Figure 7.2: The top panel shows model $F_s$ (solid line) and $F_v$ (dash line). The panels below show modelled (lines) and measured (dots) (b) snow depth (c) latent and (d) sensible heat fluxes at GB2 in 2003. Snow depth measurements were manual and flux measurements were automatic.
Figure 7.3: Same as Figure 7.2 but for P2 in 2003. Unlike in Figure 7.2, modelled snow depth (b) are against automatic snow measurements.
Figure 7.4: Same as Figure 7.2 but for GB2 in 2004.
Figure 7.5: Same as Figure 7.2 but for P2 in 2004.
Figure 7.6: Modelled (black lines) latent heat fluxes at the bare ground and snow surfaces and sensible heat fluxes at the ground, snow and vegetation surfaces at GB2 in 2003.
Figure 7.7: As Figure 7.6 but for P2 in 2003.
7.4 Comparison between 3SM and JULES simulations

In order to quantify the need for a 3-source model for use in shrub-tundra environments, point simulations were also performed with JULES at GB2 and P2. Results are shown in Figures 7.8 and 7.9 for one year only (2004) because differences between the two schemes are consistent between years.

The top plot in both figures shows that point runs in JULES are unable to deal with the heterogeneity of the surface. The snow cover fraction is calculated in JULES following Equation 3.14, which is graphically presented in Figure 3.2, which causes $F_s$ to decrease quickly on the assumption that, at a point, snow is either present or not. On the other hand, the model structure in 3SM allows it to represent horizontal heterogeneity and, therefore, the snow cover depletion curve is less steep.

The canopy model in JULES causes both the latent and sensible heat fluxes to be overestimated. As JULES is unable to deal with horizontal processes, the model considers the surface to either be snow or canopy, unlike 3SM which is able to simulate both at the same time. As a consequence, when the canopy model is switched on in JULES, as is needed when running it over shrub-tundra, JULES calculates turbulent fluxes from the canopy. If no snow is diagnosed on the canopy the heat exchange is assumed to be between canopy and the atmosphere. The low albedo of the canopy causes it to have a high surface temperature and therefore sensible heat fluxes are overestimated because heat exchanges between the snow to the atmosphere is neglected. With regards to $LE$, following Equation 3.40, the moisture flux neglects sublimation from snow on the ground when the canopy model is switched on. Nevertheless, the latent heat fluxes are higher in JULES than in 3SM because of the evapotranspiration from the canopy.

When snow-free, heat fluxes modelled with JULES are closer to observation than heat fluxes modelled with 3SM. As was mentioned in the previous section, it is possible that the overestimation of the heat fluxes in 3SM is due to the absence of a soil model. It will therefore be essential to address this issue if 3SM is to be used to model the annual energy balance of shrub-tundra.

Nevertheless, Figures 7.8 and 7.9 confirm what was explained in Section 6.4.1.1, namely that JULES was developed to deal with vertical processes but that the model structure is not currently able to deal with horizontal processes, which are dominant in sparse vegetation landscapes. A multiple source model is therefore more adequate to investigate snow melt energetics in shrub-tundra.
Figure 7.8: Modelled snow cover fraction and observed (red dots) snow depth, latent heat and sensible heat fluxes against model simulations with 3SM (black line) and with JULES (green line) at GB2 in 2004.
Figure 7.9: As Figure 7.8 but for P2.
7.5 Distributing the driving data

The Granger basin, where the distributed model is to be run, has slopes in excess of 20° but an altitudinal range of less than 260 m. As a consequence, it was deemed necessary to distribute the incoming shortwave radiation and the wind speed but not air temperatures or precipitation. The processes used to distribute the two meteorological components are described below.

7.5.1 Distribution of the incoming shortwave radiation

The solar geometry calculations described in this section are taken from Oke (1987) and Liston and Elder (2006).

\( \hat{S} \), the solar radiation per unit surface area, depends upon the local slope and solar geometry and the partitioning between diffuse and direct radiation such that

\[
\hat{S} = \left( \frac{S_{\text{dir}} \cos \hat{\Theta}}{\cos Z} + S_{\text{dif}} \right),
\]

(7.23)

where \( S_{\text{dir}} \) and \( S_{\text{dif}} \) are the direct and diffuse radiation on a horizontal plane respectively, \( Z \) is the solar zenith angle and \( \cos \hat{\Theta} \) is the angle of incidence between the normal to the slope and the solar beam. \( \cos Z \) is calculated as

\[
\cos Z = \sin \Phi \sin \delta + \cos \Phi \cos \delta \cos h,
\]

(7.24)

where \( \Phi \) is the latitude of the site, \( \delta \) is the solar declination and \( h \) is the hour angle calculated as

\[
h = \frac{\pi}{12} (12 - t),
\]

(7.25)

where \( t \) is the local solar time. \( \delta \) is approximated as

\[
\delta = -\Phi_T \cos \left( 2\pi \frac{\text{DOY} + 10}{365} \right),
\]

(7.26)

where \( \Phi_T \) is the latitude of the Tropic of Cancer and DOY is the day of the year. The slope is related to solar geometry in \( \cos \hat{\Theta} \) such that

\[
\cos \hat{\Theta} = \cos \beta \cos Z + \sin \beta \sin Z \cos (\Omega - \hat{\Omega}),
\]

(7.27)

where \( \beta \) is the slope angle, \( \hat{\Omega} \) is the slope azimuth angle and \( \Omega \) is the solar azimuth angle measured clockwise from north and calculated as
\[ \Omega = \arccos \left( \frac{\sin \delta \cos \Phi - \cos \delta \sin \Phi \cos h}{\sin Z} \right) \]  \hspace{1cm} (7.28)

if \( t < 12 \) and

\[ \Omega = 2\pi - \arccos \left( \frac{\sin \delta \cos \Phi - \cos \delta \sin \Phi \cos h}{\sin Z} \right) \]  \hspace{1cm} (7.29)

if \( t > 12 \).

Given a digital topographic map, the slope angle can be calculated as

\[ \beta = \tan^{-1} \left[ \left( \frac{\partial z}{\partial x} \right)^2 + \left( \frac{\partial z}{\partial y} \right)^2 \right]^{1/2}, \]  \hspace{1cm} (7.30)

where \( z \) is the topographic height and \( x \) and \( y \) are the horizontal coordinates, and the slope azimuth angle as

\[ \hat{\Omega} = \frac{3\pi}{2} - \tan^{-1} \left( \frac{\partial z/\partial y}{\partial z/\partial x} \right). \]  \hspace{1cm} (7.31)

Following the empirical method of Erbs et al. (1982), atmospheric transmissivity, \( k_t \), is

\[ k_t = \frac{SW_\downarrow}{I_0 \cos Z}, \]  \hspace{1cm} (7.32)

where \( I_0 \) is the solar constant. The diffuse radiation is then \( S_{\text{dif}} = dSW_\downarrow \) where for \( k_t > 0.8 \), the diffuse fraction \( d = 0.165 \), for \( k_t \geq 0.22 \)

\[ d = 1 - 0.09k_t \]  \hspace{1cm} (7.33)

and for \( 0.22 < k_t < 0.8 \)

\[ d = 0.9511 - 0.1604k_t + 4.388k_t^2 - 16.638k_t^3 + 12.336k_t^4. \]  \hspace{1cm} (7.34)

Direct radiation is then simply \( S_{\text{dir}} = S - S_{\text{dif}} \). A map of distributed incoming shortwave radiation on a clear sunny day is presented in Figure 7.10.
7.5.2 Distribution of the wind component

A simple wind model written by Richard Essery, University of Edinburgh, was used to distribute wind speeds relative to those recorded at P2. Following the wind flow scheme in Liston and Sturm (1998), wind speed was increased on windward slopes and ridges and decreased on lee slopes and in hollows. Slope over distance $s_{upwind}$ of a point, $s$, is

\[ s \propto z(0) - z(l_s) \]  

(7.35)

where $z(0)$ is the point elevation. $s$ is positive on windward slopes and negative on lee slopes. Curvature along the wind direction over distance $l_c$ is

\[ c \propto 2z(0) - z(l_c) - z(-l_c) \]  

(7.36)

which is positive on ridges and negative in hollows. The model for wind speed ratio relative to P2 is given by

\[ u = a_0 + a_s s + a_c c \]  

(7.37)
where the coefficients $a_0$, $a_s$ and $a_c$ are given by multiple regression with wind speed maps produced by the Mason and Sykes model, which calculates wind flow over topography (Mason and Sykes, 1978). For south-westerly winds over GB, adjusting $l_s$ and $l_c$ to maximise the correlation gives $l_s = 1583 \text{ m}$, $l_c = 79 \text{ m}$, $a_0 = 1.14$, $a_s = 0.0025 \text{ m}^{-1}$ and $a_c = 0.028 \text{ m}^{-1}$. A map of distributed wind speed at a timestep on a windy night (23 April 2004), calculated by using the normalised wind fields produced with the model, is shown in Figure 7.11.

### 7.6 Distributed Initial conditions

#### 7.6.1 Snow depth and SWE

The Distributed Blowing Snow Model (DBSM), details of which can be found in Essery et al. (1999) and Essery and Pomeroy (2004b), was used to generate initial snow conditions at GB. The model predicts snow cover characteristics taking into consideration variations in topography and vegetation characteristics (vegetation fraction, stem density and canopy height).
DBSM is driven by meteorological data (air temperature, relative humidity, wind speed, wind direction and snowfall) and informed by a DEM, a canopy height map and a fractional vegetation cover map. Changes in SWE within a gridbox over time are calculated as

\[
\frac{\partial SWE}{\partial t} = S_f - q_s - \nabla . q_t
\]  

(7.38)

where \( S_f \) is the snowfall rate, \( q_s \) is the sublimation rate and \( \nabla q_t \) is the divergence of snow transport in and out of the gridbox.

DBSM was run from 1 October 2003 to 15 April 2004 on a 8 m x 8 m grid. Figure 7.12 shows modelled snow depth obtained at the end of the run in the 1 km x 1 km portion of GB. Figure 7.13 shows the same snow depth against manual measurements along the three snow survey transects at GB (Figure 2.7).

Although the model tends to underestimate snow depth, it reproduces the distribution of snow well. It is able to reproduce topographically-driven snow vari-
Figure 7.13: Initial snow depth from DBSM (line) against manual measurements (crosses) along the A, C and F transects. The grey bands show the range of snow depth within 8 m on either side of each transect.
ability such as the location of the snow drift situated at the top of the north-facing slope, situated 20 to 50 m into the transects (see Figure 2.7 for elevation and topography along the transects). Shrub-driven snow cover variability is also simulated on both slopes and in the valley. Figure 7.14 shows modelled snow distribution against LiDAR canopy height. Snow depth on the north-facing slope are relatively close to shrub height, showing that the model is able to simulate the trapping of snow by shrubs. Equally, at the end of the south-facing slope along the F transect, two deep snow patches (320 m - 350 m and 360 - 400 m respectively) occur within 5 m downwind of shrub patches which act as snow fences.

As snow distribution in shrub tundra depends upon topography and vegetation cover, the performance of DBSM depends highly upon the quality of the vegetation and topographic maps used to inform the model. As was seen in Section 2.2, there are still issues with LiDAR-based canopy height measurements for short vegetation and manual measurements sometimes differ by up to 1 m from LiDAR measurements. A better vegetation map may improve the performance of DBSM and provide more accurate initial conditions for D3SM.

In order to accommodate the boundary conditions of the model, results from the distributed runs only show the central 1 km x 1km portion of the Granger basin.
Figure 7.14: Initial snow depth from DBSM (line) against LiDAR-based vegetation height (triangle) along the transects.
7.6.2 Soil moisture factor

The soil moisture factor $\psi$ was distributed by assuming a linear relationship between the topographic index and soil moisture. These assumptions form the basic concepts of TOPMODEL, which are described in Section 4.8. A simple version of TOPMODEL, available on the University of Lancaster Hydrology and Fluid Dynamics Group’s website (Hydrology and Fluid Dynamics Group, Lancaster University, 2005), was used to calculate a topographic index map of the basin (Figure 7.15). The map clearly identifies the stream channels, which are denoted by high $\chi$ values. $\psi$ was then distributed such that

$$
\psi = \psi_{P2} + \frac{\chi - \chi_{P2}}{\chi_{GB2} - \chi_{P2}} (\psi_{GB2} - \psi_{P2})
$$

(7.39)

where $\psi_{P2}$ and $\chi_{P2}$ are the soil moisture factor and the topographic index at P2 and $\psi_{GB2}$ and $\chi_{GB2}$ are the same respectively at GB2.

Figure 7.15: Topographic index over the Granger basin.
7.7 D3SM results

In the absence of continuous meteorological station data leading to Spring 2003, D3SM was evaluated against snow depth and turbulent fluxes measurements from Spring 2004 only (see Section 2.2.1 for details). Figures 7.16 and 7.17 show modelled snow depth against manual measurements on 2 days during the melt season in 2004. On 6 May, modelled snow depth are closer to observations on the north-facing slope than on the south-facing slope, where, unlike the model, manual measurements already show a number of snow-free patches. On 16 May, D3SM manages to closely reproduce the location and depth of all the late-lying snow along the three transects. It is unclear whether the model is creating a late lying snow patch 170 m into the A transect at the end of the run or if this disagreement between model and observations is due to field sampling errors. Indeed, modelled and measured snow depth at the same distance matched closely on 6 May and none of the other points with similar snow depth on that day are snow-free in Figure 7.17. On the other hand, the model late-lying snow patch situated from 340 to 380 m into the F transect is more likely due to errors with the initial conditions as snow depth were already overestimated at the start of the run. As was mentioned in Section 7.6.1, errors with pre-melt snow distributions may be caused by the quality of the data (such as vegetation characteristics) used by DBSM. However, investigating the cause of errors in DBSM is beyond the scope of this study.

Distributed snow cover in the basin is shown in Figure 7.18, where the three transects are marked as in Figure 2.3. Although the transects are still snow-covered, the more exposed south-facing slopes in the northeast of the basin are already snow-free. Smaller exposed rock outcrops where little snow gather in the winter are also snow-free on the north-facing slope.

On the last timestep of the model run (Figure 7.19) the largest area of snow remaining on the ground is situated on the north-facing side of the valley which is also downwind of the predominant wind direction (Figure 2.9). On the plateau north and south of the valley late-lying patches due to topographic hollows remain although snow cover is greater on the north facing side of the basin.

The influence of snow distribution in the basin is reflected in the distribution of turbulent fluxes averaged over the duration of the run (Figures 7.20 and 7.21). Sensible heat fluxes are highest over areas that became snow-free early in the model run or have exposed vegetation. On the contrary, the largest negative latent heat fluxes are on the north-facing slope and south of the valley, where
deep snow still remains at the end of the run.

7.8 Summary

A model calculating separate energy balances for each sub-area within a gridbox, namely vegetation (shrub), snow and bare ground was introduced. Point runs established that the model was able to reproduce the snow depth and the heat fluxes over short and tall shrub sites. The distributed model was also found to capture the snow distribution, the evolution of the snowpack and the heat fluxes over a heterogeneous landscape. Although the model was found to perform well against snow depth during the snowmelt season along three 400 m transects, further work into evaluating model snow cover distribution against photographs or remote sensing products is required. Nonetheless, D3SM was found to be an efficient tool to calculate snowmelt energetics. It will therefore be used in the next chapter to investigate the possible effects of shrub expansion on snow and heat fluxes in a shrub-tundra landscape.
Figure 7.16: Modelled against measured snow depth on 6 May 2004 along the A, C and F transects.
Figure 7.17: Modelled against measured snow depth on 16 May 2004 along the A, C and F transects.
Figure 7.18: Modelled snow cover on 6 May 2004 over GB. White pixels are snow and brown pixels are snow-free. The three black lines are the A, C and F transects, as shown in Figure 2.3.

Figure 7.19: Same as Figure 7.18 but for 16 May 2004.
Figure 7.20: Mean sensible heat flux over the duration of the run at GB. The geographic coordinates are as in Figure 7.18. The arrows point to the area covered with tall shrubs in the valley.

Figure 7.21: Mean latent heat flux over the duration of the run at GB. The geographic coordinates are as in Figure 7.18.
Chapter 8

Effect of shrub expansion on snow, soil temperatures and the energy balance in a sub-Arctic basin
8.1 Introduction

Since Chapin et al. (1995) showed that climate warming favoured shrub growth in Arctic tundra, many studies have investigated the factors affecting energy exchange and snowmelt in shrub-tundra landscapes (e.g. Sturm et al., 2005b; Pomeroy et al., 2006; Bewley et al., 2007, 2010; Marsh et al., 2010). However, most studies have also considered the effect of shrub expansion on snowmelt energetics by assuming that the vegetation shift would mostly occur from grassland to shrubland. Following Tape et al. (2006), who suggested that shrubs colonised bare ground through the expansion of existing shrub patches, in this chapter, the potential effect of shrub expansion on the energy exchange, snowmelt and soil temperatures at sites where shrubs are already in the landscape is investigated.

8.2 Control of topography and shrub cover on pre-melt snow distribution

Before investigating the effect that shrub expansion may have on a shrub-tundra site (Granger basin), the effect that the current shrub distribution has on the current snow distribution and energy exchange need to be quantified. In order to facilitate this objective, the blowing snow model, presented in Section 7.6.1, provided the pre-melt snow conditions over GB for a D3SM run in which the vegetation was removed. A second run was performed with the current shrub cover but without topography in order to differentiate the influence of shrubs from the influence of landforms on snow distribution in the model grid. The initial SWE from these runs and from the run used in the previous chapter (hereafter referred to as the control run) are presented in Figure 8.1. As with the runs performed in the previous chapter, results here are shown only for the central 1 km x 1 km portion of the basin.

The correlation coefficient of the SWE between the control run and the no-shrub run is higher ($r^2 = 0.87$) than between the control and the no-topography run ($r^2 = 0.48$), suggesting that topography has a greater influence on the SWE in the basin than shrub cover. This pattern can also be seen on the 3 transects, where $r^2$ for snow depth between the control and the no-shrub run equals 0.78, 0.92 and 0.88 along the A, C and F transects respectively, against $r^2 = 0.61$, 0.48 and 0.65 between the control and the no-topography run (Figure 8.2).

Peaks in snow depth in the shrub-only run correspond to peaks in shrub height (Figure 8.2). However snow depth under each peak is not proportional to shrub
height. For example, on the A transect the deepest snow (0.98 m) is found under a 1.16 m tall shrub whereas snow depth under the tallest shrub (2.42 m) is only 0.88 m deep. This is consistent with Sturm et al. (2001a) and Essery and Pomeroy (2004b) who reported a threshold in shrub height (∼1 m) above which snow holding capacity does not increase with increasing shrub height. Although the correlation coefficients of the $SWE$ between the no-topography and the control runs are weaker, they still indicate a modest correlation, which shows that shrubs do influence snow distribution, although to a lesser degree than the topography of the basin does. Furthermore it is likely that topographical gradients contribute to the spatial variability of shrub cover itself. The correlation coefficient between the LiDAR-derived vegetation fraction (Figure 2.3-b) and the topographic index (Figure 7.15) indicate that 43% of the spatial variance in shrubs is correlated to topography and, by extension, to soil moisture and nutrient availability. The greener areas in the vegetation map clearly coincide with the highest indices (e.g. low slope and valley bottoms) whereas the bare areas coincide with the lowest indices (e.g. high slope and hill tops). This finding agrees with Ostendorf and Reynolds (1993) who calculated exactly the same correlation coefficient between patch-scale NDVI and topography in an Alaskan catchment. In the following sections, a series of runs were performed with D3SM (one without topography, one without shrubs, and two with increased vegetation cover and canopy height) in order to determine if an increase in shrub cover will change the snowmelt energetics in the basin and reduce the influence of topography on snowmelt energetics in GB.
Figure 8.1: Initial snow water equivalent in (a) the no-shrub (b) the no-topography (c) the control runs.
Figure 8.2: Initial snow depth along the A, C and F transects in the control (black), the no-topography (red line) and the no-shrub (green) runs against shrub height (dots).
### 8.3 Increasing shrub cover in the Granger basin

In order to investigate the influence of shrub expansion on snowmelt energetics with D3SM, modified vegetation cover maps are needed. In the absence of an ecological model to simulate shrub growth, two methods, each intended to represent ways in which shrub cover may change, were used to “grow” shrubs in as-yet uncolonized patches and to densify the already existing shrub patches.

1. As shrub expansion occurs mainly around existing patches (Sturm et al., 2001b; Tape et al., 2006), vegetation was grown on and near cells with existing cover. For cell \(n\) with \(F_v = F_{vn}\), \(F_v\) in each neighbouring cell (8 in total) was read. If at least one of the neighbouring cells had \(F_v > F_{vn}\), \(F_{vn}\) was increased by a random number between its own value and the largest \(F_v\) value of neighbouring cells. The whole process was repeated twice. The resulting \(F_v\) map is shown Figure 8.3-a. This expansion scenario is hereafter referred to as D3SM-Patch.

2. Following the relationship between topography, soil moisture and vegetation cover discussed in the previous section and in Section 4.8 and 7.6.2, shrubs were expanded as a function of the topographic index, \(\chi\). First \(\chi\) and \(F_v\) for cell \(n\) were determined. Then \(F_{vn}\) was increased to a random number between \(F_{vn}\) and the maximum \(F_v\) in the next \(\chi\) increment. This method aims to simulate shrub expansion solely as a function of increased soil moisture availability, as could happen if some of the permafrost underlying GB thaws in response to increasing air temperatures. The resulting \(F_v\) map is shown Figure 8.3-b. This expansion scenario will hereafter be referred to as D3SM-TOP.

\(F_v\) increases from a basin average of 0.08 in the control run to 0.17 in D3SM-Patch and 0.41 in D3SM-TOP. Canopy height was increased following the same method in both scenarios: for each cell where \(F_v\) was increased, canopy height was randomly increased by no more than a third (average \(h_c\) in D3SM-Patch is 0.89 m against 0.86 m in D3SM-TOP). Although this value is nominal, it aims to take into consideration environmental conditions that are known to limit shrub growth but that are not accounted for in the modified vegetation runs, like abrasive wind on the exposed plateau or nutrient availability in the riparian zones.

In order to assess how reasonable these shrub expansion scenarios are, the modified \(F_v\) in D3SM-TOP and D3SM-Patch are compared with results from Tape et al. (2006). They found that the percentage of ground covered by shrubs increased on average by 33% from 1945 to 2002 in Northern Alaska, with a specific
Figure 8.3: Maps of the modified vegetation fraction in (a) D3SM-Patch (b) D3SM-TOP.
36% increase on slopes. Although their assessment was retrospective, \textit{i.e.} of shrub expansion that had already happened, they also predict that the rate of expansion is increasing. In this study, 61\%, 97\% and 89\% of the pixels were covered by shrubs in CTL, D3SM-Patch and D3SM-TOP respectively, showing that D3SM-Patch incurred a 36\% increase and D3SM-TOP a 28\%. Both increases are therefore within the range of values reported by Tape et al. (2006).

8.3.1 Comparison of pre-melt snow depth between the control and the modified runs

In order to facilitate the discussion below, two dominant features in the basin are distinguished in Figure 8.4: shrubs taller than 1.2 m and snow deeper than 0.8 m at the beginning of the run. The arrows point to the snow drift on the north-facing slope.

Figure 8.5 shows the difference in initial snow depth between the control run and the modified runs (positive values indicate deeper snow in the modified runs). Figure 8.5a corroborates the strong influence of topography on snow distribution in the basin. Removing the topography leads to the disappearance of the large snow drift and other deep snow patches which correspond to those in Figure 8.4. These patch-scale drifts occur in the control run because they form in topographic
depressions in the lee of the predominant south-westerly winds (see Figure 2.9 for wind directions in GB). In the absence of hollows for snow deposition in the flat terrain run, more snow is available for trapping by shrubs. This is clearly seen from the band of deeper snow which roughly corresponds to the band of tall shrubs in Figure 8.4.

On the other hand, removing shrubs, and therefore a snow trapping mechanism, causes snow to be shallower in that same area (Figure 8.5b). Shallower snow depths in the no-shrub run roughly correspond to areas of shrub cover in Figure 2.4. This is particularly noticeable on the right side of the transects where average $F_v = 0.27$ (against 0.08 for the whole basin). Unlike in Figure 8.5a, deep snow is now found in topographic hollows where, as no snow is being trapped upwind by shrubs, more snow is available for deposition.

Differences in snow depths in Figures 8.5-c and 8.5-d show that the greater the basin average $F_v$ is, the more homogeneous snow depth is. This is quantified in the standard deviation (STDEV) of SWE, which decreases with denser $F_v$ (Table 8.1). Unlike in Figures 8.5a and 8.5b, there are less large scale reorganisations of snow distribution. Instead the changes occur on much smaller patch-scales, as can be seen by many blue pixels being adjacent to red pixels, most probably because of the patchy nature of the changed $F_v$.

As in the no-topography run, D3SM-TOP (Figure 8.5d) shows shallower snow in the snow drift at the top of the north-facing slope and in topographic hollows. However, the change in snow depth is also very patchy and many areas where snow depth increased are adjacent to areas where snow depth decreased. This occurs because more snow is trapped by shrubs upwind therefore less is available for redistribution to topographically-driven drifts. However, the correlation coefficient of SWE between D3SM-TOP and the no-topography run is still lower ($r^2 = 0.42$) than between D3SM-TOP and the no-shrub run ($r^2 = 0.70$), which means that topography still dominates the spatial variability of snow in the basin. However, the strength of the correlation between the no-shrub run and the vegetation runs, and therefore the influence of topography on snow distribution, does decrease with increasing $F_v$. Because of the way the two shrub expansion scenarios were designed, an increase in $F_v$ is defined here as an increase in both densification of existing shrubs patches and pixels covered by shrubs. Although, as seen in Section 8.3, the shrubs cover a larger fraction of the basin in D3SM-Patch (0.97), where $F_v > 0$, they are less dense in the former (average $F_v = 0.17$) than in the latter (average $F_v = 0.41$).
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<th></th>
<th>( F_v )</th>
<th>initial ( SWE ) (mm)</th>
<th>STDEV initial ( SWE ) (mm)</th>
<th>final ( SWE ) (mm)</th>
<th>( H ) (W m(^{-2}))</th>
<th>( LE ) (W m(^{-2}))</th>
</tr>
</thead>
<tbody>
<tr>
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<td>123</td>
<td>95</td>
<td>44</td>
<td>-4</td>
<td>4</td>
</tr>
<tr>
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<td>122</td>
<td>100</td>
<td>44</td>
<td>-8</td>
<td>11</td>
</tr>
<tr>
<td>No-topography</td>
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<td>116</td>
<td>77</td>
<td>25</td>
<td>4</td>
<td>5</td>
</tr>
<tr>
<td>D3SM-Patch</td>
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<td>123</td>
<td>89</td>
<td>37</td>
<td>-1</td>
<td>4</td>
</tr>
<tr>
<td>D3SM-TOP</td>
<td>0.41</td>
<td>124</td>
<td>80</td>
<td>35</td>
<td>3</td>
<td>5</td>
</tr>
</tbody>
</table>

Table 8.1: Initial \( SWE \), standard deviation of initial and final \( SWE \) across the basin and one month basin average sensible and latent heat fluxes.
Figure 8.5: Difference in snow depth at the beginning of the runs between the control run and (a) the no-topography run (b) the no-shrub run (c) D3SM-Patch (d) D3SM-TOP. The black lines are the A, C and F transects.
8.3.2 Snowmelt energetics with increasing vegetation

The basin average snowmelt rates, $SWE$, $F_s$, $F_v$, sensible and latent heat fluxes for the duration of the run are shown Figure 8.6. Differences in latent heat and sensible heat flux between the control run and the modified topography and vegetation runs, averaged over one month, are shown in Figures 8.8 and 8.9 respectively. The one-month average $H$ and $LE$ from the control run were shown in the previous chapter, in Figures 7.20 and 7.21 respectively.

As a general rule, the snowmelt energetics and processes in the flat plane runs show rather different patterns from the four other runs. This is not surprising as topography was determined to be the dominant process controlling snow depth in the basin.

The no-topography run has the lowest initial basin-average $SWE$ but the highest $F_s$. The low $SWE$ is due to the absence of the snow drift marked in Figure 8.4. Although no quantitative assessment is available for the whole basin, the drift was found to store up to 65% of the snow water equivalent along the three transects (Quinton et al., 2004). Without this wind-sheltered, north-facing drift, the snow is redistributed over the whole basin and is more exposed to sublimation and melt. The high $F_s$ occurs because snow can lie where slopes are normally too steep to allow snow cover. The spatial homogeneity of the snow cover remains throughout melt, as seen in Figure 8.7a. Except along the tall shrub line where pixels become snow-free from days 123 to 131, the rest of the basin is either still snow-covered or has become snow-free within the last 4 days of the run. The no-topography run is also the one that has the greatest snowmelt in one month, losing 78% of its snow water equivalent, because the whole basin receives the same amount of radiation. Snowmelt rates are the highest of the runs from days 118 to 128 but the lowest towards the end of the run. This transition from highest to lowest melt rates coincides with two events: an increase in sensible heat fluxes and a decrease in $F_s$. Between day 127 and 129, $F_s$ goes from being the highest amongst all runs to being the lowest. Following this the available energy is transformed predominantly into sensible heat from the shrubs rather than into melt, starting on the first day that $F_v + (1 - F_s)(1 - F_v) > F_s$, or in other words, when the snow-free fraction becomes greater than the snow cover fraction.

The no-shrub run produces more than twice as much latent heat as the other runs (Figure 8.6, Figure 8.8, Table 8.1). This occurs because evaporation from the shrub tile is assumed to be negligible (see Section 7.2.2) and thus removing shrubs from the basin allows for a larger area from which $LE$ can be calculated. Conversely, sensible heat fluxes are the smallest amongst all runs (Figure 8.9b).
because positive $H$ can only come from the snow-free ground rather than from the snow-free ground and the shrubs in the other runs.

The catchment average $F_s$ in the no-shrub run is the lowest throughout the month, with the exception of the no-topography run from day 128 onwards (Figure 8.6). It is also the one that shows snow-free pixels the earliest in the season (Figure 8.7b). On the other hand, the no-shrub basin average $SWE$ is consistently the highest alongside that of the control run. As seen in Table 8.1, the standard deviation of $SWE$ at the beginning of the run in the no-shrub run is the highest of all the runs, but the average $SWE$ is close to all but the no-topography. These values suggest that, in the absence of vegetation, snow cover is more heterogeneous, as can be seen in Figure 8.1a where snow is deeper than in Figure 8.1c in drifts and topographic hollows but shallower in the valley. However, they also suggest that snowmelt is slower. As was seen in Figure 7.6, the vegetation surface produces large sensible heat fluxes. Therefore, in the absence of shrubs, there is no advection of sensible heat from the shrubs to the snowpack and snowmelt is slower than in the other runs.

The small differences in $LE$ between the vegetation runs with topography (control, D3SM-Patch and D3SM-TOP) in Figure 8.8c-d are almost entirely due to changes in $LE_q$. Out of all the runs, rocky outcrops on the south-facing slope become snow-free earliest in the control run, therefore more latent heat is produced from the ground tile than in the other two runs. However, the most significant difference between these runs is in the sensible heat fluxes. $H$ decreases in a few gridboxes (Figure 8.9b-d), where $F_v = 0$ in the control run but $F_v > 0$ in the increased vegetation runs, because the extra snow that is now trapped by shrubs causes larger negative $H_s$ fluxes. However, while all runs have the same one-month basin average $LE$ to within 1 W m$^{-2}$ (Table 8.1), $H$ increases with increasing $F_v$. Of most significance for the land surface - atmosphere interactions is the positive one-month average sensible heat flux with D3SM-TOP (3 W m$^{-2}$), which denotes a change in the flux direction compared to the less vegetated runs.

Chapin et al. (2005) calculated that the observed shortening of the snow season in spring, of 2.5 day per decade on average, increases the energy transferred from the land surface to the atmosphere by 26 MJ m$^{-2}$ year$^{-1}$ i.e. 3.3 W m$^{-2}$, which is comparable in terms of atmospheric heating to a doubling of atmospheric CO$_2$. The results presented in this study suggest that shrub expansion, could cause increases in heat transfer to the atmosphere twice as large as those suggested in Chapin et al. (2005).

In Section 7.3.2, 3SM was found to predict sensible heat fluxes almost twice as
large as measurements once the ground was snow-free, which could suggest that
the positive $H$ with D3SM-TOP is due to model errors. Furthermore, the one-
month catchment average $H$ with D3SM-TOP is just 3 W m$^{-2}$ which means that
accounting for the overestimation of sensible heat in the model may decrease the
average and revert to negative $H$.

In an attempt to quantify the error in $H$, the one-month average of measured and
modelled sensible heat fluxes was calculated at the two meteorological stations in
GB (see Figure 7.2 to Figure 7.5 for measurements and model results). Out of the
four datasets, the D3SM sensible heat flux average was larger than measurements
twice, once at GB2 in 2003 (66 W m$^{-2}$ with D3SM against 52 W m$^{-2}$) and once
at P2 in 2004 (14 W m$^{-2}$ with D3SM against 5 W m$^{-2}$) and lower at GB2 in
2004 (22 W m$^{-2}$ with D3SM against 36 W m$^{-2}$) and at P2 in 2003 (43 W m$^{-2}$
with D3SM against 52 W m$^{-2}$). However, averaging sensible heat fluxes over
both sites and both years shows that the model average $H$ is exactly the same as
the measured average (36 W m$^{-2}$). As a consequence, it is likely that the model
is able to capture well the increase in sensible heat with increasing vegetation
and that the estimated difference of 7 W m$^{-2}$ between CTL and D3SM-TOP is
reliable.

An equation of the form $H = a \ln(F_v) + b$ was fitted to the data ($r^2 = 0.99$) in
order to obtain an estimate of the critical shrub cover at which sensible heat would
become positive in GB. The results suggest that the basin will become a heat
source for $F_v \geq 0.22$. Further work into assessing the rate of shrub expansion in
the Granger basin may determine the timescale at which this shift could happen.

### 8.3.3 Summary of the results

Investigations using D3SM into snow distribution in the Granger basin showed
that topography exerts a stronger control on the spatial variability of snow than
shrub distribution. The vegetation fraction and canopy height were increased to
assess the effect of shrub expansion on snowmelt energetics. While topography
was still found to have a large influence, making shrubs denser and expanding
shrub patches reduced the spatial variability of snow depth and increased the
snow cover fraction. D3SM also found that increasing the vegetation fraction
increased sensible heat fluxes, which are, as a consequence, predominantly from
the surface to the atmosphere during snowmelt.
Figure 8.6: Time series of area average snowmelt, $SWE$, $F_v$, $F_s$, sensible and latent heat fluxes for the control, no-shrub, no-topography, D3SM-TOP and D3SM-Patch runs.
Figure 8.7: First snow-free day over GB in (a) the no-topography run (b) the no-shrub run (c) the control run (d) D3SM-Patch and (e) D3SM-TOP. A white or light grey pixel means that snow had not fully melted by the end of the run.
Figure 8.8: Difference in latent heat fluxes between the control run and (a) the no-topography run (b) the no-shrub run (c) D3SM-Patch (d) D3SM-TOP.
Figure 8.9: Difference in sensible heat fluxes between the control run and (a) the no-topography run (b) the no-shrub run (c) D3SM-Patch (d) D3SM-TOP.
8.4 Effects of air temperature and snow depth on soil temperatures

Previous researchers have reported that a shift from open- to shrub-tundra would lead to deeper snow because wind-blown snow lost to sublimation in open environments remains in the landscape in shrub-tundra (Sturm et al., 2001a; Liston et al., 2002; Essery and Pomeroy, 2004b). However, the results in the previous section showed that increasing shrub cover in shrub-tundra did not automatically increase snow depth, but rather reduced the spatial variability of snow. As was discussed in Section 1.4, the structure and depth of the snowpack has a strong influence on subnivean temperatures. Snow provides low conductivity to the underlying surface and therefore exerts a strong control on shrub growth by promoting microbial activity and nutrient availability. In addition, it regulates the temperature and depth of the active layer where large amount of methane, unaccounted for in the carbon cycle, are stored (Fukuda, 1994; Michaelson et al., 1996; Anisimov et al., 1997). It is therefore essential to assess how shrub-snow interactions could change the soil thermal regime.

In Section 3.7.4, soil temperatures at one shrub site in the WCRB (the buck brush site or BB) and at one nearby alpine site (AS) were found to considerably differ. The elevation at which measured soil temperatures at the alpine site were taken coincides with the maximum elevation at GB. Equally, the elevation at which measured soil temperatures at the buck brush site were taken coincide with the minimum elevation at GB. In the absence of a year long soil temperature dataset for GB, soil temperatures at these two sites are used as proxies for what could happen to soil temperatures in GB should shrub expansion lead to changes in patch-scale snow depths. Descriptions of the sites and of the differences between the two sites are in Section 2.2.2.

In the absence of a soil model in D3SM, JULES, which was found to simulate measured shallow soil temperatures well in Section 3.7.4, is used to perform the controlled experiments.

8.4.1 Model experiment set-up

Two simple sets of model experiments, with modified driving data, were performed in order to compare the effect of air temperature and snow depth on soil temperature.

The original driving data were modified such that snow depth predicted at one site in the modified run are similar to those predicted in the original of the other. As
higher wind-induced compaction of the snowpack and higher sublimation rates at AS require more snowfall than in BB for equal snow depth, snowfall was decreased by a factor of 1.7 at BB to simulate AS snow depth but increased only by a factor of 1.5 at AS (top panels in Figure 8.10). The experiments are hereafter referred to as $dS_f+$ for snowfall increase in AS and $dS_f-$ for snowfall reduction in BB.

A second set of two runs was performed with air temperatures increased by 2°C at the alpine site (hereafter referred to as T2°C+) and decreased by the same amount at BB (hereafter referred to as T2°C-). The value of 2°C was chosen because a rise in air temperature of 2°C above preindustrial level is the equilibrium temperature target above which impacts associated with climate change are expected to increase markedly (Solomon et al., 2007).

The soil temperature simulation with the multiple layer snow model in Section 3.7.4 constitute the control run referred to as CTL. In the modified snowfall experiment at BB, snow depth between the first snowfall and the 19 November at is < 0.1 m, which means that the composite snow / soil scheme (CL) is used to calculate soil temperatures. As was seen in Section 3.7.4, CL underestimates soil temperatures and therefore, the results for all runs during that period are not considered in the discussion below or in Table 8.2.

8.4.2 Results

At the alpine site, winter soil temperatures in $dS_f+$, where snow is deeper than in the other runs, are on average 2.7°C and 3.2°C warmer than in CTL and T2°C+ respectively. Increasing air temperatures only causes average snow depth to be 3 cm shallower than in CTL and thus makes little difference to soil temperatures. On the other hand, from mid-May to mid-June, earlier melt of the snowpack in T2°C+ leads to soil temperatures being on average 8.7°C and 9.2°C warmer than in $dS_f+$ and CTL respectively. In spite of this, annual soil temperature is still the highest in $dS_f+$ (Table 8.2), showing that snow depth is more critical to soil temperature than increased air temperature in AS. Most significantly for winter processes, the largest difference in soil temperature between CTL and $dS_f+$ occurs at the end of December, when soil temperature is 5.8°C warmer in the latter.

At BB, the coldest winter soil temperature is predicted with $dS_f-$, where average snow depth is the lowest. Higher summer soil temperature in CTL lead to the annual average soil temperature being the highest at -0.4°C, 0.8°C more than T2°C- and 1.2°C more than $dS_f-$.

At both sites, winter and annual soil temperatures are warmest under the deepest
Table 8.2: Average snow depth ($S_d$), winter and annual air ($T_a$) and soil temperature for the 3 runs in the alpine and the Buckbrush sites.

<table>
<thead>
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<th>Alpine site</th>
<th></th>
<th>Buckbrush site</th>
<th></th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>CTL</td>
<td>$dS_f^+$</td>
<td>$T_2^\circ+$</td>
<td>CTL</td>
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<tr>
<td>$S_d$</td>
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<td>0.11</td>
<td>0.25</td>
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<td>-2.1</td>
<td>-2.6</td>
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<tr>
<td>Winter $T_{soil}$</td>
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<td>-6.3</td>
<td>-9.0</td>
<td>-6.0</td>
</tr>
<tr>
<td>Annual $T_{soil}$</td>
<td>-2.3</td>
<td>-0.5</td>
<td>-1.0</td>
<td>-0.4</td>
</tr>
</tbody>
</table>

Figure 8.10: Soil temperatures and snow depths for JULES runs with original driving data, modified snowfall and modified temperature.
snowpack, \( dS_f \) in AS and T2°- in BB. Modifying air temperature makes little or no difference to winter soil temperature. However, the shortening of the snow season in T2°+ compensates for cool winter soil temperatures. As a consequence, the difference in average annual soil temperature between CTL and T2°+ is proportionally much larger than the difference in winter. Equally, T2°- in BB has mean annual soil temperature 0.8°C cooler than CTL even though no soil temperature difference occurs in winter.

Average soil temperatures in CTL at the alpine site are colder than in either of the modified runs at BB, which means that they are affected by conditions that are not modified in these experiments. Most significantly, wind speed, which is critical in the calculation of the aerodynamic resistance and, by extension, of the turbulent fluxes, was not modified at either site although it is on average 49% faster at AS than at BB. Nevertheless, the results presented here show that soil temperature at two sub-Arctic sites is more susceptible to changes in snow depth than to changes in air temperature.

### 8.4.3 Summary of the results

As seen in the previous section, shrub expansion is likely to affect snow distribution and snow depth (Sturm et al., 2005a). Here, the results suggest that a secondary effect of this expansion is changes to patch-scale soil temperature, on which 50% and 70% change in snow depth were found to have more impact than a 2°C change in air temperature would. This finding is in agreement with Lawrence and Slater (2010) who attributed up to 30% of pan-Arctic soil temperature variations to deeper snowpack. More studies on the interaction between shrub expansion, soil temperatures and snow cover are needed in order to assess the effect of shrub encroachment on soil processes. The implications of such changes are discussed in the final chapter.
The aim of this thesis, as stated in the first chapter, was to improve our understanding and model representations of snow and runoff processes in shrub-tundra. In this final chapter, we present concluding remarks on the implications of our results and suggestions for future work, followed by a final summary of the thesis.

9.1 Conclusions

The results of the investigation into the effect of shrub expansion on snowmelt energetics in shrub-tundra, presented in Chapter 8, suggest that the expansion and densification of shrub patches will positively feed back to atmospheric warming by increasing sensible heat fluxes to the atmosphere. This change in warming is predicted even though the model includes a shrub bending parameterization that allows shrubs, where appropriate, to be buried under the snowpack and thus to reduce the exposed vegetation fraction at the beginning of the snowmelt season. If increasing shrub abundance results from warmer air temperatures, as has been suggested by a number of studies (i.e. Sturm et al., 2001b; Chapin et al., 2005; Sturm et al., 2005b; Tape et al., 2006), then shrub expansion will further promote the warming of the Arctic by providing a darker and warmer land surface.

Another important feedback loop is the one linking shrub growth to soil temperatures. Shrubs trap snow, snow insulates the soil and warmer soil temperatures promote further shrub growth. Shrubs also prevent the formation of highly conductive wind-compacted layers by sheltering the underlying snowpack, causing further insulation (Liston et al., 2002). Previous studies have shown that spatial transitions from grassland to shrub-tundra increase snow depth and thus promote warmer soil temperatures. Here, results showed that expansion of existing shrub patches, in a basin whose snow distribution is mainly controlled by topography, reduces the spatial variability of the snow water equivalent and thus may decrease snow depth at patch length scales. This could lead to a negative feedback where, by reducing the spatial variability and depth of snow, shrub expansion causes cooler winter soil temperatures, reducing microbial activity and limiting further shrub growth. Therefore, predicting the extent of shrub expansion and shifts in ecosystems at high latitudes is critical to investigations into water, energy and carbon balances in a warming Arctic.

The performance of D3SM in simulating snow depth and turbulent fluxes over GB suggests that the model is an efficient tool for investigating the snowmelt energetics of shrub-tundra. The model addresses the need, which was expressed by Sturm et al. (2005b), Pomeroy et al. (2006) and Bewley et al. (2010), to account
for the bending of shrubs under the snowpack in energy balance calculations, by incorporating an exposed vegetation fraction parametrised from a shrub bending model. The model also addresses the known limitation of dual-source models in reproducing snow melt rates over a discontinuous snow cover in shrub-tundra (Bewley et al., 2010) by calculating separate energy balances for snow, bare ground and vegetation.

However, while the model has successfully addressed critical needs in modelling shrub-tundra processes, further work is needed. Firstly, D3SM needs to be evaluated at different shrub-tundra sites. This is particularly important for the applicability of the shrub bending parameterization, whose parameters were derived or calibrated using observations from a single site at GB. Secondly, before D3SM can be used to model long term changes at the land surface, it needs to be evaluated against winter processes. This could involve the coupling of D3SM to a model like the Distributed Blowing Snow Model, used in Chapters 7 and 8. Finally, vertical processes need to be incorporated by adding a snowpack and a soil model. As was seen in Sections 8.3.2 and 8.4, shrub expansion is expected to affect the energy balance, but also the soil temperature, of shrub-tundra. There is a critical need for models to represent soil processes in shrub-tundra because warmer soil temperatures could alter the thermal regime of the permafrost, potentially liberating large amounts of methane in the carbon cycle currently unaccounted for in general circulation models (see e.g. Fahnestock et al., 1998; McGuire et al.; Sturm et al., 2005a).

However, research into the impact of shrub expansion in the Pan-Arctic will remain limited until shifts in ecosystems are predicted by robust predictive vegetation models that take into consideration high latitude-specific processes. Many ecological models that simulate patterns of vegetation change have oversimplified representations of Arctic ecosystems, where factors that are known to regulate plant growth at high latitudes are often not included (Kittel et al., 2000; Sitch et al., 2007). For example, most ecological models neglect mosses, permafrost and topography, which are critical to soil moisture availability and, by extension, to shrub growth.

As was seen in Chapters 4 and 6, improvements to the representation of soil moisture heterogeneity in land surface models are also needed. The changes implemented in JULES led to improvements in streamflow simulations in two basins in Fennoscandia and to the identification of likely sources of error in the model, related to the absence of a sub-grid elevation parameterization. However, further work is needed to evaluate the transferability of the new parameterization.
to other Arctic basins and, most importantly, to ungauged basins. Warmer air temperatures are expected to shift the timing of peak streamflow from spring to winter (Parry et al., 2007). In addition, changes at the land surface, such as shifts in vegetation, will affect the snow water storage and, therefore, the volume of streamflow. Information on terrestrial water storage will become increasingly important for decision makers in order to adopt strategies related to water resources, infrastructure, flood forecasting or hydroelectric power (Furgal and Prowse, 2008; NRTEE, 2009).

While in the past models often needed to simplify representations of processes happening at the land surface, continual advances in computational efficiency are allowing models to explicitly incorporate an increasing number of processes. At the same time, intensive field campaigns are improving our understanding of the physical processes and interactions occurring at high latitudes (e.g. IP3, 2006; ABACUS, 2006). It is therefore essential that progress in understanding the interactions between atmospheric, ecological, biophysical and hydrological systems is used to improve predictive models. Of most significance is the need to relate these systems to predictive vegetation models if a number of assumptions are to be removed from impact studies on the effect of shrub-tundra expansion on the energy, water and carbon balances at high latitude.

9.2 Summary of this thesis

Investigations into snow and runoff processes in shrub-tundra were performed with the community land surface model JULES (Joint UK Land Environment Simulator), which calculates energy, water and carbon fluxes between the land and the atmosphere. After describing the science behind JULES, the performance of the model in shrub-tundra environments was evaluated against observed runoff in two basins in Fennoscandia. The model predicted peak flow too early in the snowmelt season as well as overestimating peak flow rates. The cause for the poor performance of the model was attributed to the poor representation of sub-grid soil moisture heterogeneity, leading to poor infiltration during the spring snowmelt and subsequent errors in surface and subsurface runoff partitioning. In order to address this issue, alternate formulations of soil processes were tested. The formulation which was found to perform the best in the basin that spans from boreal forest to Arctic tundra included two modifications to the original code. The first one was a modification to the hydraulic conductivity calculation so as to allow infiltration of liquid water through the frozen soil fraction. The
second modification used the relationship between soil moisture and topography to calculate a subgrid saturated fraction from the local water table depth. This formulation, which changed infiltration rates during the snowmelt season, allowed a better representation of summer streamflow. However, the model was still found to perform poorly at the shrub-tundra basin in Abisko, Sweden. We suggested that one of the causes of the model errors was that JULES does not allow shrubs to get buried in the snowpack in winter. In order to address this issue, the Shrub Bending Model (SBM), which calculates the exposed winter shrub fraction, was developed. SBM works by bending shrubs under the weight of snow and simulates a process which affects the albedo and sensible heat fluxes of shrub-tundra landscapes. SBM was evaluated against measured vegetation fractions and albedo from a shrub-tundra basin (Granger basin, GB) in the Yukon Territory, Canada, and was able to capture their progressive change over a discontinuous shrub cover. The exposed vegetation fraction obtained with the model was then parametrised for use in JULES alongside the snow cover parameterization. Allowing shrubs to be buried in winter and to become exposed progressively during spring resulted in higher spring albedo values and slower snowmelt, giving a better fit to streamflow rates.

JULES was then used to investigate the effect of shrubs on snow and runoff processes at the two sites. The current structure of the model was found to be inadequate for investigations into shrub expansion. The canopy model in JULES was developed to simulate vertical processes, such as snow interception and below canopy sensible heat and longwave radiation exchange, which are more appropriate for use in boreal forests. However, horizontal processes, such as heat advection from branches and snow-free patches, are critical in sparse canopy landscapes.

To address this issue, a surface energy balance model, the Distributed 3-Source Model (D3SM), in which separate energy balances are calculated for the shrub, bare ground and snow surfaces, was developed. Unlike JULES, which represents surface heterogeneity using a tiling scheme where different surface types interact independently with the atmosphere at a reference height, the three sources in D3SM interact with each other and are coupled to the atmosphere at the canopy height. This model structure allows energy advection from exposed vegetation and snow-free ground to contribute to snowmelt for patchy snow and vegetation cover on short length scales. The model was evaluated against snow and energy flux measurements from the Granger basin and was found to perform well. Although sensible heat fluxes over snow-free ground were overestimated, the model
was found to simulate snowmelt energetics better than JULES at the same site. It was suggested that adding a soil model to D3SM would improve its simulation of sensible heat fluxes.

Investigation into the spatial variability of snow in the basin showed that topography was the controlling factor over snow distribution. A moderate but smaller correlation between shrub distribution and topography was also found. D3SM was then used to investigate the effect of denser vegetation cover on the energy balance of the basin by increasing the vegetation fraction and canopy height of the current shrub distribution. The model predicted that shrub expansion in the basin would reduce the spatial variability of snow depth. D3SM also found that, while increasing shrub cover made little difference to latent heat fluxes, the average sensible heat flux switched from being negative, *i.e.* to the land surface, to being positive, *i.e.* to the atmosphere. This finding suggests that shrub expansion, which is promoted by warming air temperature, may positively feed back to the atmosphere by contributing to increases in local air temperature.

Experiments using JULES examined the relative importance of changes in snow depth and air temperature in terms of their impact on soil temperature. The aim of these experiments was to assess if snow depth or air temperature was the controlling factor over soil temperature at both sites. The experiments showed that a 50% change in snow depth had more effect on soil temperature than a 2°C change in air temperature. The implication for soil temperature where shrubs are expanding, is that the changes in snow depth predicted in this study could affect the thermal regime of the soil more than an increase in air temperature.

Finally, the implications of the results presented in this thesis on future modelling studies were discussed. We concluded that there was a need to relate hydrological, ecological and geomorphological spatial heterogeneity in models in order to understand complicated interactions at the land surface and make predictions.
References


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