



Technical Report No. 16

Process that impact runoff generation in Northern Latitudes



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1. Introduction

It is possible and necessary to model river flows in the current climate. Water resource engineers and natural hazard engineers require accurate forecasts of the river flows for given precipitation scenarios. A high level of accuracy is possible as the computer models can be trained on past events. But, to extend that prediction accuracy into the following decades will require more physically-based models, with less training as the boundary conditions of the problem move away from conditions we have encountered. This is particularly the case in the Northern latitudes, where the river flows are the consequences of multiple causes: some coincident, some sequential, all of them important and all of them subject to change in the new warmer climate of the Arctic and Boreal zones.

This technical report lays out the fundamental processes that affect the river flows in the North and how they are represented in both hydrological and meteorological models. The question that it aims to answer is: to what processes are the river in the northern latitudes sensitive and how can we represent these processes in our large scale models?

2. Background

Surface runoff generation occurs when water is not able to soak into the soil and is forced to run off the land either onto the neighbouring down-stream patch of land, or into the river system. The process is key to the river's hydrograph, as water that has travelled via this surface route enters the river system much quicker than water that percolates through the soil and travels to the river network via sub-surface routes. It therefore plays a large role in Hydrological Models and much research has been carried out to get this feature right for different regions. Meteorologists, who have no operational interest in the timing

of flow peaks in rivers, have invested less resources on this process in their Land Surface Models (used by Meteorologists as part of their Weather and Climate Models), whose primary function is the partition of incoming energy into sensible and latent heat.

However, in the last few years, Land Surface Models have increasingly been used as pseudo Hydrological models at the global scale: testing the ability of Global Climate models to reproduce and predict changes in the global water cycle. There is some sense in this as many processes, such as evaporation and the soil moisture control of runoff generations, are key to both hydrology and meteorology and are included in models from both communities. However, given their original purposes, the hydrological models tend to have very simple representations of the processes that affect the surface energy balance, and more complex representations of the processes that affect the water balance, and vice versa. Although the overall balances may be right at a longer time or space scale, the hydrological models will place a higher priority in getting the timing of the river flow right (dependent upon the partition of the surface to subsurface runoff) while the land-surface models will aim to get the timing of the evaporation fluxes right at the hourly timescale.

But why can't the models get everything right? Why the need to prioritise and make compromises? The reason for these compromises was different for each group. Hydrological models were hampered by lack of driving data to resolve the energy balance. Assumptions had to be made about the energy available for evaporation and snow melt, for instance. However, now there are globally and nationally available data of the sub-diurnal meteorological forcing fields for years. It is no longer necessary to guess what the evaporative demand on a particular day is. Land Surface models, on the other hand, were hampered by the need to keep their calculations to minimum. The models were used as part of larger atmospheric model and computer capacity limited the number of calculations allowed or the scale and resolution at which the models could be run. A large part of the modeller's job was to reduce the physics down to the absolute minimum. This is no longer necessary. Computer power has increased dramatically and the limits can be lifted.

So, while the history of these models created this divide, their future use is beginning to converge and this merging requires a new approach to process modelling. This new approach, whereby hydrological models and land surface models converge is a large part of the philosophy of WATCH. It is the intention to bring together the physical understanding of these two communities together to build a new generation of land surface hydrology models that combine the best of both.

3. Northern Latitudes

One region which represents a particular challenge to this merging of modelling strategies is the Northern Latitudes. The reasons are manifold as there are many processes involved. The big hydrological event of the year in these regions is the spring melt of the snow. The representation of this event requires that you have the correct timing of the snow melt. Snow melt is physically complex: first you have to know how much energy there is, which depends on the albedo, which depends on the age of the snow and how dirty it is. With very low sun angles in these regions, the amount of sun the land surface receives can vary dramatically with the slope of the land and whether it is north or south facing, or whether it is covered with trees.

Then you have to know if it melts out altogether or refreezes within the snow pack. You have to know the amount of snow you have. The amount of snow that falls is hard to measure (observational data of snowfall is notoriously uncertain) but will almost certainly vary with altitude. The amount of snow on the ground at the time of melt may not be the amount that fell there, since it both blows about and

evaporates during the winter. It can settle on tree tops, or fall to ground below depending on what type of tree you have.

If you know the snow melt, you then have to know how much of the water will infiltrate into the soil. This is an unusually difficult problem in northern latitudes as the soils are very spatially varied, with patches of organic soils next to mineral soils. The depth of the soils and the permeability of the substrate also vary across a landscape: some soils are underlain by rocks, some are very deep while others lie on permafrost. In addition, the soil may be frozen, maybe partially frozen or may be patchily frozen. The soil may be frozen at some depth for part of the year or all year. In addition, the northern latitudes are typically very wet with the water table close to the surface and distinct and extensive areas of wetlands and lakes.

All of these processes are unique to Northern latitude landscapes and all of them have a big impact on runoff generation. What is also true of this landscape is that it is remote and often inhospitable. Observations are therefore difficult to obtain and therefore relatively rare. Both hydrological and land surface modellers of these regions have had to deal with far greater uncertainty than other regions because of the difficulty of obtaining some of the most essential data, such as snow fall.

The hydrological models of this region tend to cope with the complex nature of the processes and the uncertainty of the data by taking the key outputs that are needed, and finding empirical equations that robustly reproduce these outputs. This has taken an enormous amount of work (as evidenced by the number of papers on the subject, only a fraction of which are reported here) and represents a huge resource for future researchers. The land surface models of this region typically take a different approach and put as much physics into the model as possible with the aim that, at least, changes are then robustly represented.

This report aims to describe in some detail the processes that affect the runoff generation in northern latitudes (outlined above) and describe what some land-mark hydrological models do (in particular the Cold Regions Hydrology Model, CHRM: www.usask.ca/hydrology/crhm.htm), what land-mark land surface models do (in particular CLM, the Community Land Model: www.cgd.ucar.edu/tss/clm/distribution/clm3.0 and the JULES mode: www.jchmr.org/jules being developed under this project). The report will then describe a new way forward, combing the best of both. Each process is covered in a separate section. Each section will identify short, medium and long term goals for modelling the runoff generation in northern latitudes.

4. Description of each process with current and possible future parameterisations

The first five sections describe snow processes. The first is about the importance of the lateral heterogeneity of snow cover: how it gets like that and what models aim to do about it (section 4.1). The next four discuss science of the point process that affect snow melt, including the vertical distribution of snow in a tree-ground combination (section 4.2), the basic physics of radiation (4.3), sublimation (4.4) and snow melt (4.5).

4.1 Snow Distribution – lateral heterogeneity of snow patches

The overall energy balance of a snow-covered landscape is strongly affected by the fractional coverage of the snow. The dark patches of ground between the snow packs absorb ten times more radiation than

the snow patches. There had been a theory that the no-snow patches directly heated the snow patches through a lateral transfer of sensible heat. Several research studies aimed at quantifying this process concluded that the lateral transfer was negligible, although the darker no-snow patches tend to warm up the area and thereby accelerate the snow melt. In addition, at very small scales, such as occurs when the vegetation is taller than the snow pack and therefore sits in the same one-meter-scale space, there is likely to be some long-wave radiation transfer from the warm vegetation to the snow.

In the current version of JULES (v2.0), it is only possible to include the effect of the warming of the air temperature by the non-snow patches or the transmission of long-wave radiation from the non-snow covered vegetation to the snow, indirectly through an effective reduction of the albedo. This is the default setting for the small-scale case described above: for a given snow depth, different vegetations can be either buried by the snow or can poke out of the snow pack, depending on their height. The fraction of the grid box assumed to be covered by snow, F , is given by:

$$F = \frac{s}{s + A_1 z_0} \quad (1)$$

where s is the snow depth (m), A_1 is a factor (in this case it is 10 which is the standard conversion from roughness length to vegetation height) and z_0 is the roughness length of the vegetation (1/10 of the vegetation height).

Some other methods have been proposed in the literature (e.g. Roesch et al. 2001, Essery and Pomeroy 2004). For instance, it has been proposed that the effect of mountains on the fractional snow cover (due to aspect and roughness) can be included as follows:

$$F = \frac{s}{s + A_2 \Omega} \quad (2)$$

where A_2 is a factor (in this case 0.15, taken from Roesch et al. 2001) and Ω is the Standard Deviation of the orography (m).

Another new method is proposed which takes account of the additional impact of within-grid temperature changes with altitude. The only available information on the sub-grid topography for this case study is the standard deviation of altitude. If it can be assumed that the altitude varies linearly between plus and minus 1 standard deviation, and that the temperature varies with altitude according to the adiabatic lapse rate (0.6 degrees for every 100m), then the fractional cover of snow can be described as follows:

$$F = 0.5 + 0.5 * \left(\frac{(T - T_m) * 100}{0.6 * \Omega} \right) \quad (3)$$

where T is the air temperature (K) at the mean altitude of the grid box and T_m is the melt temperature (273.15 K) and Ω is the Standard Deviation of the orography (m).

Each of these formulae (Equations 1, 2 and 3) are tested separately within the JULES model in a simulation of Northern Europe, all using the same model-generated driving data for current climate, to quantify any differences to the modelled snow cover.

The impact of this on the mean snow cover in May for the twenty years of simulation is shown in Figures 1a-c. The default method of calculating snow-cover (Equation 1) gives the lowest snow cover fractions, Equation 2 (which account for the roughness of mountains) makes a difference in the mountainous regions of the Urals. The method that accounts for the effect of altitude (Equation 3) gives a strong north-south gradient to the snow cover (reflecting the cooler temperatures in the north).

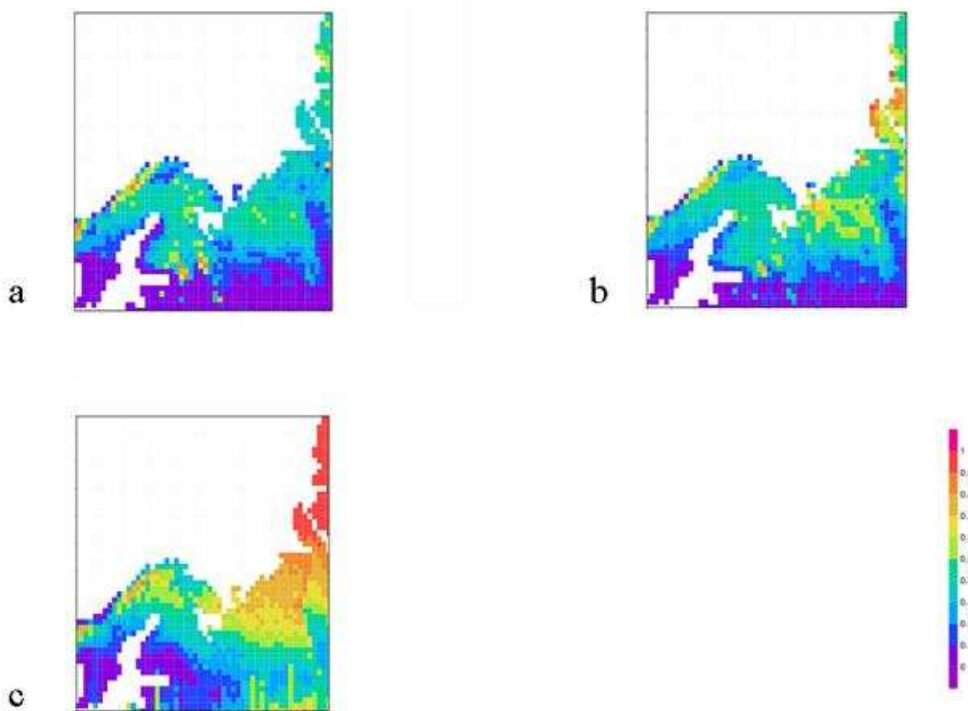


Figure 1a. Snow cover for May for current climate as predicted by JULES, using Eqn 1
 Figure 1b. Snow cover for May for current climate as predicted by JULES, using Eqn 2
 Figure 1c. Snow cover for May for current climate as predicted by JULES, using Eqn 3

These studies demonstrate the importance of the fractional snow cover on the overall landscape energy balance, snow melt and therefore runoff. However, it is not satisfactory to include this important process using a proxy, i.e. the effective albedo. Indeed, the lack of an explicit fractional snow cover in JULES has been a cause of some concern which led to some research by Wiltshire (2006) who demonstrated the importance on runoff of including separate energy flux calculations for snow and no-snow patches within a single JULES grid. Ideally, this indirect method would not have to be used. The results are summarised in Figure 2.

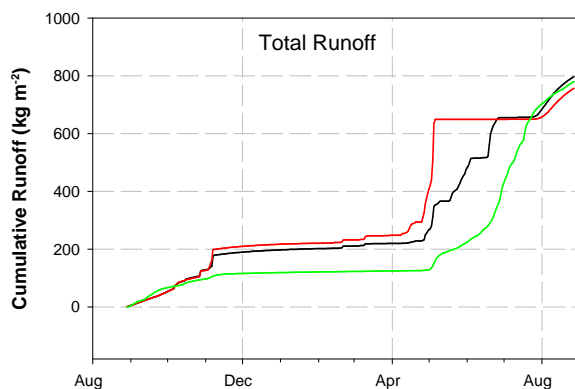


Figure 2: Simulated catchment runoff from the separate energy balance scheme (black) and single energy balance scheme (red) against observations from a stream gauge (green).

The logical conclusion from this research for modelling is to allow the snow patches to have a separate energy and radiation balance. This is now included in the JULES model. It is yet to be fully tested and a short term goal is to test this model against available data (observed and model derived snow heterogeneity). It is likely that the small-scale impact of vegetation protruding through the snow pack on albedo may be kept in the JULES model, although a more thorough treatment of the long-wave interaction between vegetation and snow using a two-source model would be a long-term goal. This solution is the subject of current research at CEH and Edinburgh University.

Once a separate energy balance is possible, the next stage is to quantify the actual distribution of snow depths (including the fraction of no-snow) across landscape at any one time. A study which tested whether the snow melt of variable snow depths occurred at the same rate was carried out by Essery et al (2005) using data from a site in Svalbard. They demonstrated that, if the radiation forcing was the uniform, all the snow melted at the same rate. This means that to calculate the overall snow melt, all that is needed is statistical descriptions of the initial snow depth across a landscape and the variable atmospheric drivers and a good snow model will dictate the rest. For instance, as lateral heat transfer is not an issue, it is not necessary to account for the patch size.

So a big research question is therefore, what is the distribution of snow before the melt takes place? There are many factors determining whether snow stays present in an area, and determining its distribution within the landscape, once it has fallen. As well as the impact of large-scale orography on precipitation (precipitation may increase with elevation and decrease on lee slopes), topography affects the spatial distribution of snow in several ways: by affording shelter to wind-blown snow, by affecting melt through altering the angle of the ground to the incoming radiation and by the effect of altitude which reduces air temperature with height. These effects have been studied at several different scales:

- A) At the micro-scale (~100m) the dominant process is the shelter effect: Blowing snow drifts behind obstacles or in depressions and these snowdrifts can store a considerable proportion of the snow; Pomeroy et al. (1997), for example, estimated that drifts covered 8% of the area but held 30% of the snow at maximum accumulation for a 68 km² basin with moderate topography in the Canadian Arctic. The presence of clumped vegetation also has this effect so that large snow-drifts accumulate at the edges of forests and in areas of shrubs (Pomeroy et al. 1993, Sturm et al. 2001).
- B) At meso-scales (~1km) the impact of the shelter is also important and strong winds can sweep a whole hill-side clean of snow, while the relatively sheltered areas will accumulate the snow. In addition, at this scale the aspect of the surface is important, so that north-facing slopes retain their snow cover for much longer into the spring than the south-facing slopes. Pomeroy et al. (2003) showed how this influenced the snowcover on north- and south-facing slopes of a valley in the Yukon, Canada.
- C) At the landscape scale (~10km) and above, the effect of altitude on temperature from adiabatic cooling of the atmosphere, has an effect on how long the snow lasts. Observational studies at this scale have not yet been published in the literature, but the effect is evident to any lay observer, and some large-scale models divide mountainous grid cells into elevation bands to account for it (Arola and Lettenmaier 1996, Liston et al. 1999, Essery 2003).

The blowing-snow process has been the subject of much research, which has resulted in some practical models, in particular a two-dimensional snow cover mass balance (Pomeroy *et al.*, 1993) which was later extended to three dimensions (Essery *et al.*, 1999). This model is used in CRHM, modified to a single column calculation with new methods to calculate the inputs and to scale the fluxes from a point

to a landscape in an areal snow mass balance calculation (Pomeroy *et al.*, 1997). The following, up-scaled, mass balance can be drawn over a fetch with distance, x (m),

$$\frac{dSWE}{dt}(x) = P - p \left[\nabla F(x) + \frac{\int E_B(x) dx}{x} \right] - E - M$$

where $dSWE/dt$ is surface snow accumulation rate ($\text{kg m}^{-2} \text{s}^{-1}$), P is snowfall rate ($\text{kg m}^{-2} \text{s}^{-1}$), p is the probability of blowing snow occurrence within the fetch, F is the downwind transport rate ($\text{kg m}^{-2} \text{s}^{-1}$), E is snow surface sublimation rate ($\text{kg m}^{-2} \text{s}^{-1}$), E_B is the blowing snow sublimation rate ($\text{kg m}^{-2} \text{s}^{-1}$), and M is snow melt rate ($\text{kg m}^{-2} \text{s}^{-1}$). Essery and Pomeroy (2004) have recently shown that a simple system of sources and sinks based upon vegetation height can provide similar basin average snow mass to a fully distributed blowing snow model.

For the purposes of Land Surface modeling, which works at scale from 1km up to 100km, we necessarily consider this to be a sub-grid process and aim to parameterize the process by assuming that snowfall is immediately deposited into two snow-buckets: one deep one shallow. The shallow bucket melts out fast and the deep bucket stays longer. The key question is to find the buckets sizes both in terms of depth and extent. A medium term goal would be to parameterize this two-bucket model using data at different scales to understand the link with vegetation distribution, micro and macro topography and prevailing meteorological conditions.

A longer term goal is to include a flexible snow-patch tile, where the tiles include deep snow and shallow snow to represent the redistribution of snow as it falls, altitude tiles to represent the differential precipitation and melt rates due to temperature and aspect tiles to represent the different radiation loading of different parts of the landscape. As part of this longer term goal, we would wish to identify ways to populate these tiles. At the 10km scale, the parameterizations can be checked by comparing the output to a fine-scale distributed model run. At the larger scale, EO data of snow cover will be used.

4.2 Snow Distribution – vertical

A crucial component of snow covered landscape is the extent to which the snow fall to the ground below the canopy of the trees. It has an impact not only on the overall albedo of the landscape, with the dark trees absorbing 10 times more energy than if the snow were exposed to the sky. But also the snow below typically lasts longer into the spring, being subject to low winds, low aerodynamic roughness and reduced radiation. Observations of these effects have been the subject of much study in the Yukon (Sicart *et al.* 2004, Bewley *et al.* 2006), Alaska (Sturm *et al.* 2005) and Colorado (Hardy *et al.* 2004).

This effect was demonstrated to be important to land surface models, when a pronounced improvement to forecasts of the ECMWF model was traced back to the year that this process was included. The JULES model includes such a vertical distribution and has been shown to have a big impact on the runoff generation (Essery and Clark, 2003).

An optional canopy / snow interaction model (canopy model 4) was implemented in JULES in which the snow interception at the branch scale is applied to the whole canopy. Snowfall is partitioned into interception by the canopy and throughfall to the ground. Sublimation can occur from the intercepted snow but the formulation of the moisture flux neglects sublimation from snow on the ground, thus leading to a deeper snow pack than with the previous snow scheme. Figure 3 shows the difference in modelled daily climatological snow water equivalent (SWE) and Figure 4 shows the subsequent runoff

in the Abisko catchment between using this new version (can_mod_4) and the original where the snow was kept up on the top of the vegetation (can_mod_3).

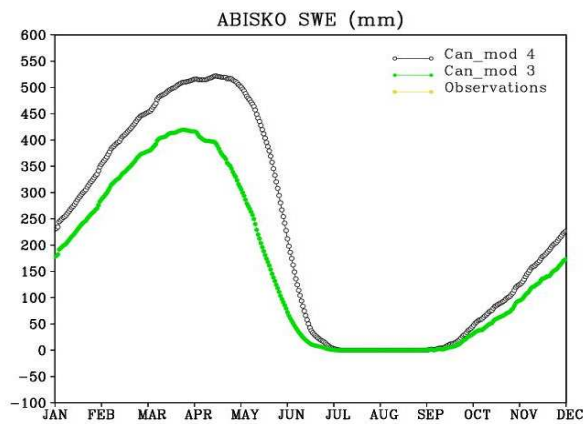


Figure 3: Modelled daily climatological snow water equivalent in Abisko, Sweden

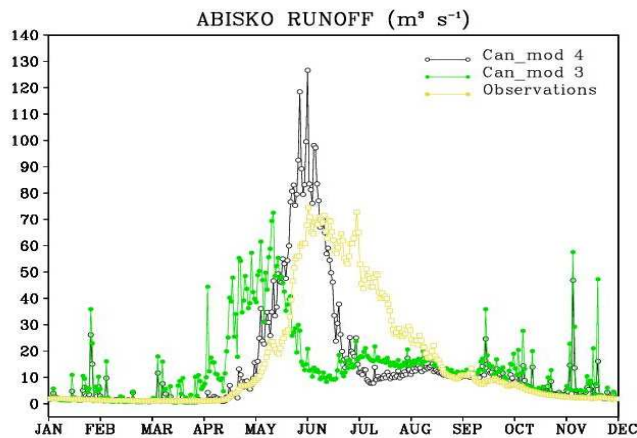


Figure 4: Modelled and observed daily climatological runoff in Abisko, Sweden.

Clearly, this is an important process for runoff generation. A short term goal is to check the performance of this feature against large scale river flow data in regions away from the Abisko region where it has been tested thoroughly.

A particular tree type in northern latitudes is the sparse birch. This is an extensive biome. To quantify the radiative properties of sparse trees, and how much sunlight is reaching the snow on the ground, requires a complex three-dimensional, time-dependent approach, since at low sun angles the sun only reaches the trees, whereas at high sun angles, the trees occupy only about 10% of the area. Earth Observation scientists have built models of tree canopies to solve this physical problem. A similar approach is being adopted by the land surface modellers: research into the radiation balance of sparse trees is being carried out with field work, leading to an improved version of JULES. A medium term goal is to include the feature of below-canopy snow with deciduous trees. A medium term goal is to include the dynamic sun-angle model of radiation in JULES.

4.3 Radiation balance of snow and heterogeneity

There is a reduction in snow albedo due to snow aging (assumed to represent increasing grain size and dirt, soot content) resulting in earlier snowmelt. The CRHM model addresses this issue by splitting the melt regimes into different periods or types as follows:

Premelt - from 1 Feb. up to the start of active melt albedo decreases at a relatively constant rate, except for event-caused increases due to snowfall and decreases due to melting. Rates of depletion range from 0.004 to 0.009/day with an average of 0.0061/day.

Melt - the general shape of the albedo-depletion curve during continuous melt is "S"-shaped in which the period of rapid decrease in albedo is preceded and followed by 1 to 2 days when the rate of change is slower. The decrease during rapid, continuous melt is approximated by the expression:

$$A(t) = A_i - 0.071t$$

in which $A(t)$ is the albedo after " t "- days of continuous depletion and A_i is the albedo of the snow surface at the start of "active" melt. The period of ablation of shallow arctic and prairie snowcovers under continuous melting often spans only 4 to 7 days.

Postmelt - following the disappearance of the seasonal snowcover the albedo of the ground surface takes on a relatively-constant value of 0.17 (value can be adjusted). The decrease in albedo of late-occurring snows occurs at a rate of about 0.20/day

The Snow albedo in JULES also varies with time. The aging of snow is characterized by introducing a prognostic grain size, $r(t)$, set to $r_0 = 50 \mu\text{m}$ for fresh snow and limited to a maximum value of $2000 \mu\text{m}$. The change in $r(t)$ over a timestep, t , is given by

$$r(t + \Delta t) = \left[r(t)^2 + \frac{G_r}{\pi} \Delta t \right] - [r(t) - r_0] \frac{S_f \Delta t}{d_0}$$

where S_f is the snowfall rate during the timestep and d_0 , the mass of fresh snow required to refresh the albedo, is set to 2.5 kg m^{-2} . The empirical grain area growth rate is

$$G_r = 0.6 \mu\text{m}^2 \text{ s}^{-1} \text{ for } T = T_m$$

$$G_r = 0.06 \mu\text{m}^2 \text{ s}^{-1} \text{ for } T < T_m, r < 150 \mu\text{m}$$

$$G_r = A \exp(-E / RT) \text{ for } T < T_m, r > 150 \mu\text{m}$$

The inclusion of dust and black carbon is not yet parameterized in JULES. It would be possible to do so using the same procedure as the snow-grain size module.

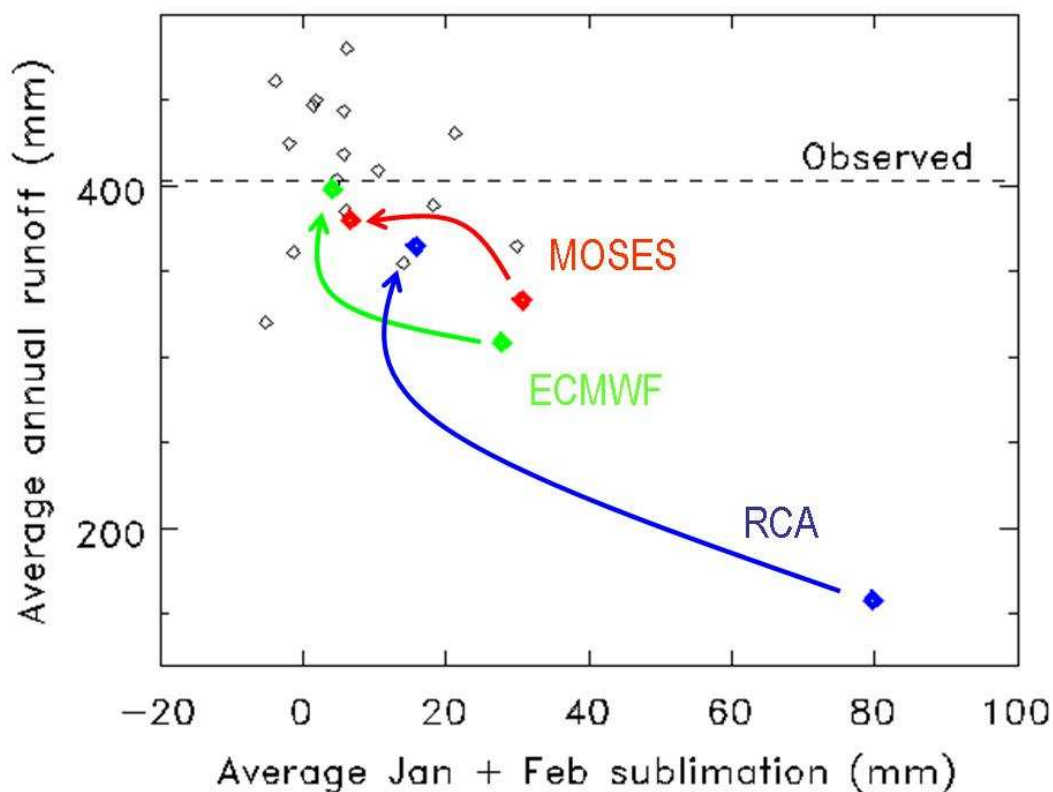
A short term goal would be to check whether JULES reproduces the calibrated CHRM. A long term goal would be to include dust/black carbon.

4.4 Evaporation/Sublimation

The process of sublimation is the same as for evaporation and from saturated parts of the surface (lakes, wet vegetation canopies and snow) is calculated at the potential rate (i.e. subject to an aerodynamic resistance only). This makes the sublimation rate very sensitive to external controls. Studies that compared model output for this were surprising in their diversity. The PILPS2e

intercomparison study (Bowling et al. 2003) demonstrated that among-model differences in winter latent heat due to the treatment of aerodynamic resistance appear to be at least as important as those attributable to the treatment of canopy interception. Even though the largest evaporation rates occur in the summer (June, July and August), model-predicted snow sublimation in winter has proportionately more influence on differences in annual runoff volume among the models.

A plot of annual runoff against winter sublimation for the 21 participating models (see Figure 4), shows that the more snow you sublimate, the less you have to melt and runoff - and the changes that three models made to their surface formulations based on this experience (the coloured arrows)



4.5 Snow Melt

Once the energy has been correctly absorbed by the snow, the next process is to heat and melt the snow. Firstly, the heating of the snow requires the snow to transmit the heat down through the pack. The important process in this is the layered nature of the snow densities. The density depends on the temperature of the snow, and the weight above it and how long it has been squashed by the snow above.

In addition, when the top layer of snow melts, in practice it does not necessarily run straight into the soil. If there are layers of snow below at lower temperatures, the water often refreezes. This ice-layer has a high density and affects the snow dynamic.

With the new layered snow model in JULES, it is possible to resolve the different density of the different layers.

Once the snow has reached zero degrees Celsius, the heat is used to melt the snow. The Latent Heat of Fusion is high and keeps the snow temperature at a constant until the snow has melted.

A short term goal is to test the new layered snow module against data.

4.6 Surface and Subsurface Runoff

Surface runoff is calculated in the time step in which there is a positive flux of water to the surface (either snow melt or rainfall). The runoff is calculated as the flux of water multiplied by the fraction of the grid box that is assumed to be saturated. There are two possible ways of calculating this fraction, one based on statistical distribution of the topographic index in the catchment and the other based on a standard statistical distribution used for hydrological modelling. Both these are options in the JULES model.

A medium term goal is to check whether observed distributions of saturated soils across an Arctic Hillside (Sayer, 2007) looks like either of these formulae. This will be useful in the formulation of runoff generation at the small scale. At larger scale, other data will be used such as wetland fraction from EO.

Additional surface water is generated if the flux is greater than the maximum infiltration rate allowed for that soil type. At present the maximum infiltration rate is determined only by the saturated hydraulic conductivity and an infiltration enhancement factor that depends on the vegetation type. The main issue here is the hydraulic conductivity of frozen soils. See Section 4.8.

Subsurface runoff is calculated as the drainage out of the bottom soil depth:

$$K = K_s \left(\frac{\theta}{\theta_s} \right)^{2b+3}$$

4.7 Vertical Soil Water Dynamics of organic soils

Linking all these processes, and therefore vital to the runoff generation, is how easily the water flows through the column of soil. For the arctic, the key issue is the presence of organic soils.

Hydraulic properties of organic soils are distinctly different from the mineral soils. They drain quickly where the density of the soil is low, for instance near the surface. But with depth, the organic matter gets compacted and the drainage reduces, sometimes to zero, leaving the soils saturated and water logged. It is important for land surface models used in areas of extensive organic soils to model this phenomenon correctly.

In an extensive study of the issue, Letts et al, 2000 demonstrated that the basic hydraulic assumptions used in typical land surface models (layered soils using the Darcy-Richard equations to calculate the vertical fluxes of water between layers) still hold for organic soils, but that the parameters need to be carefully chosen. One of the problems in obtaining soil parameters is characterising the type of organic soil in the area. While mineral soils have recognised pedo-transfer functions which can be used to obtain hydraulic and thermal properties according the composition of the soil in terms of percentage sand, silt and clay, organic soils have yet to be characterised in such a way. The percentage carbon content is one factor in determining the hydraulics of the soil, but also the density, the compaction and the type of organic content. Letts et al (2000) brought together a range of studies of observed hydraulic properties to demonstrate the issue of characterising the type of organic soil.

According to Letts et al. (2000), different classes of organic soil, for instance, Fibric, Hemic and Sapric soils, each have different hydraulic properties. For instance, Letts *et al.* (2000) shows that the range of hydraulic conductivity values can vary from 2.2×10^{-3} (ms^{-1}) (Fibric peat) to 1×10^{-12} (ms^{-1}) (Sapric peat). Bellisario *et al.* (2000) suggest values of hydraulic conductivity for use in the CLASS for each peat type (1.7×10^{-4} Fibric, 2.0×10^{-6} Hemic and 1.0×10^{-7} Sapric).

Most models to date have tended to use a shallow soil column (3m in JULES) which severely limits their ability to represent the thermal inertia associated with deep permafrost layers. This missing representation also introduces errors in the simulation of the seasonal cycle of soil temperatures (Alexeev et al., 2007, Nicolsky et al., 2007). Further, a deeper modelled soil column allows the land to store much more heat (Stevens et al., 2007); the associated increased flux of heat into the soil will tend to reduce the energy available for near-surface melting. Lawrence et al. (2008) show that these deficiencies can be readily addressed by first defining a deeper soil column and second taking account of the influence of soil organic matter on the thermal properties of the near-surface layer. The representation of permafrost in JULES is the subject of ongoing work in the development of the Quest Earth System Model (QESM).

A short term goal is to include any new parameterizations, conditioned by observations, in the version of JULES used in WATCH. A medium term goal is to test the importance of soil depth to runoff generation. A long term goal is to include some parameterization of soil depth heterogeneity in JULES.

4.9 Infiltration in frozen soils

The presence of frozen soils in the Arctic and Boreal regions results in dramatic changes in the springtime runoff peaks. This is because of the reduced hydraulic conductivity when the soils are frozen or even partially frozen.

In the CRHM, the algorithms that calculate frozen soil infiltration during snowmelt and over-winter soil moisture changes are based upon 15 years of study of the snow hydrology of the Prairie region of Canada (Gray *et al.*, 1986) and results reported in the former USSR (Motovilov, 1978, 1979; Popov, 1973). The Division of Hydrology at the University of Saskatchewan (Granger et al. 1984) postulated that the infiltration potential of frozen soils may be grouped in three broad categories, namely: restricted, limited and unlimited.

Restricted - Infiltration is impeded by an impermeable layer, such as an ice lens on the soil surface or within the soil close to the surface. For all practical purposes, the amount of meltwater infiltration can be assumed to be negligible and that the melt goes directly to runoff and a little to evaporation. *Limited* - Infiltration is governed primarily by the snow-cover water equivalent and the frozen water content of the top 30 cm. of soil.

Unlimited - A soil with a high percentage of large, air-filled macropores at time of melt. Examples of soils having these properties are dry, heavily cracked clays and dry, coarse sands. All meltwater infiltrates these soils and runoff from overland flow is negligible. Granger et al. (1984) made field observations in Bad Lake and surrounding farmland of infiltration from snowmelt to medium to fine-textured, uncracked frozen prairie soils in which entry of meltwater is not impeded by ice layers (limited case). The findings show that:

- i) the mean depth of infiltration during snowmelt was 260 mm,
- ii) infiltration was relatively independent of soil texture and land use,
- iii) the amount of snowmelt infiltration was inversely related to the average moisture content of the 0-30 cm depth soil layer (θ_p) at the time of melt.

These findings are supported by further observations and physical modelling in the Prince Albert Model Forest and Wolf Creek Research Basin boreal and tundra soils (Zhao and Gray, 1999; Gray *et al.*, 2001; McCartney *et al.*, 2006).

Significant amounts of liquid water may coexist with ice at freezing temperature and with the introduction of the concept of supercooled soil water, liquid water can exist at temperatures below freezing. The relationship between unfrozen soil water concentration and temperature when ice is present is described in terms of the soil water suction

$$\psi = -k \left[\frac{\rho_i L_f}{\rho_w T_m g} \right] (T - T_m)$$

where T_m (K) is the freezing point of pure water at atmospheric pressure, g ($m\ s^{-2}$) is the acceleration due to gravity and k is a dimensionless constant that depends on the soil. The k constant is a measure of the degree to which the adsorption of the soil dominates over the capillarity and is used as a correction factor. This typically equals 1 for clay rich soils and 2.2 for granular soils. However, for organic soils, the value of the constant is uncertain. A short term goal is to test this value against the available data.

In addition, to the presence of frozen and unfrozen soil co-existing, the other issue for the water balance is the formulation of the hydraulic conductivity. In JULES this is calculated strictly as a function of unfrozen water content such that:

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

where K and K_{sat} are the hydraulic conductivity and the saturated hydraulic conductivity, θ_u is the volumetric unfrozen water content and b is an empirical exponent.

Niu and Yang (2006) modified the formulation of the hydraulic conductivity in the Community Land Model version 2.0 (CLM2.0) in order to represent the permeability of frozen soils and the subgrid variability of the Arctic landscape in large scale modelling, such that

$$K = (1 - F_{frz}) K_{sat} (\theta_u / \theta_{sat})^{2b+3}$$

where

$$F_{frz} = \exp(-\alpha(1 - \theta_f / \theta_{sat})) - \exp(-\alpha)$$

θ_f being the volumetric frozen water content, F_{frz} being the fractional impermeable frozen area and α being an adjustable scale-dependent parameter.

When tested in JULES 2.0, this formulation of the hydraulic conductivity allowed for more water to infiltrate in the soil and to be stored in the landscape leading to delayed and less peaky runoff than with the present scheme. A short term goal is to check the performance of JULES against the CRHM model. A medium term goal is to use regional scale river flow to test the performance of the model.

5. Conclusions

Several short, medium and long term goals have been specified on this document. The following actions will further these goals:

Improved parameterizations of the processes are beginning to be included in the LSM. Each process needs to be assessed independently before confidence can be gained. It should be possible to use the CHRM model components and process level data to test the individual processes.

Heterogeneity is a key factor in runoff generation in Northern latitudes. Studies which include heterogeneity are: PILPS2e, RhoneAgg, STEPPS and CLASSIC. The results of such studies will be used to test the assumptions in the aggregation strategies. In addition, new EO datasets and large river basins will be obtained to check the large scale performance, using results of the WATCH model runs using the model intercomparison framework.

Runoff generation is impacted by a combination of processes. It is difficult to locate a key process by looking at riverflow alone. A combined data stream of riverflow, evaporation, surface temperature and snow cover would better constrain the models. Benchmark datasets that test the models will be developed for this purpose.

New parameterisations of snow heterogeneity are likely to be necessary. It is proposed that buckets and tiles will be used to represent the fractionally covered snowy landscape. Any new developments will be tested against the bench mark datasets.

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