Atmosphere/surface interactions in the ECMWF model at high latitudes

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1. Introduction

The surface boundary condition is an essential aspect of an atmospheric model as it controls the surface fluxes of momentum, heat and moisture into the atmosphere. The ocean represents a relatively simple boundary condition with a slowly varying sea surface temperature which can be kept constant to a reasonable degree of approximation in short and medium range forecasts. Although there are uncertainties in air/sea transfer over the ocean, current formulations are probably accurate within 20% (see e.g. Beljaars 1997), which is very good compared to the situation over land.

Land surfaces and sea ice are much more complex because they are highly interactive and they respond to the atmospheric and solar forcing at very short time scales e.g. resulting in a strong diurnal cycle. Furthermore, typical land surfaces tend to be inhomogeneous, which has strong impact on momentum, heat and moisture transfer. This paper focuses on the thermal aspects of the atmosphere/surface coupling over land, snow and sea ice with emphasis on high latitudes. The schemes that control the surface/atmosphere interaction in the ECMWF model have evolved enormously over the last 15 years. The land surface scheme has gradually been changed from a 2-layer model with a climatological deep soil boundary condition to a fully prognostic 4-layer model with 6 surface tiles and data assimilation for soil moisture and temperature (Viterbo and Beljaars 1995; Van den Hurk et al. 2000; Douville et al. 2000). The sea ice model was changed from a single slab model to a 4-layer sea ice temperature model and separate tiles for sea ice and open water.

In this paper a few model changes are presented which were particularly relevant to high latitude atmosphere/surface coupling. Emphasis is put on the thermal coupling including sea ice and snow. The purpose is to illustrate the relevance of the various physical aspects. The starting point is the 4-layer land surface model introduced by Viterbo and Beljaars (1995) in August 1993. It was developed and tested in single column mode using various data sets. The scheme had rather good summer performance, but fully coupled to the atmosphere it turned out that the amplitude of the diurnal cycle of temperature was too large and that the winter temperatures over continental areas were too low. These problems could be reduced by introducing the process of soil moisture freezing and by re-tuning the stable boundary layer formulation. These two aspects of the thermal coupling between atmosphere and surface will be discussed in section 2.

The BOREAS experiment was a major effort to study various aspects of the atmosphere / land surface interaction in boreal forest. One of the immediate messages from the field experiment was that snow impact on albedo was far less in forest areas than in open terrain. It resulted in a preliminary fix for snow albedo in forest areas with very positive impact on temperature biases in spring. Results will be shown in section 3.

The snow albedo, the handling of partial snow cover and terrain heterogeneity could be handled in a more consistent way with a tiled version of the land surface scheme called TESSEL. It was particularly beneficial in areas with snow, because the selected tile structure explicitly distinguishes between exposed snow and snow under high vegetation. These aspects are discussed in section 3. The tile structure also allowed for partial ice cover over the ocean and a more responsive sea ice temperature model, which is discussed in section 4.

The resulting model was used for the 45 year long re-analysis (known as ERA-40; Uppala et al. 2005) covering the period from 1957 to 2002. Section 5 discusses systematic data assimilation increments of soil moisture, soil temperature and snow water equivalent. The observations are mainly from SYNOP messages which provide indirect information on soil quantities through the model. Study of increments is interesting, because it provides information on model deficiencies. Further evaluation of the realism of the atmosphere to land surface coupling is given in section 6 making use of the BERMS data.

2. Soil moisture freezing and stable boundary layer diffusion

Before going into the effects of soil moisture freezing and boundary layer diffusion, a short description is given of the main characteristics of an earlier version of the ECMWF land surface scheme (introduced in 1994; Viterbo and Beljaars 1995). The scheme has 4 soil layers with prognostic variables for soil moisture and temperature (Fig. 1). The heat exchange in the soil is described by a diffusion equation and the vertical water movement uses the Richards equation for volumetric soil water content (Hillel 1982). Thermal soil properties and soil hydrological properties are selected for a globally uniform loamy soil type. The thermal conductivity has a weak dependence on soil moisture whereas soil hydrological properties are nonlinearly dependent on soil water content.

These diffusion type equations are solved numerically with a crude vertical discretization with 4 layers and implicit time stepping for stability. The 4 layers and their thickness are selected such that the entire spectrum of time scales from diurnal to annual can be handled. To account for the very fast time scale a skin layer is added. It has no inertia and responds instantaneously to the various terms in the surface energy balance. The skin layer represents the vegetation canopy in vegetated areas, a dirt or litter layer on top of bare soil or the top of the snow deck in snow conditions. It also intercepts the solar radiation and absorbs and emits the thermal radiation.

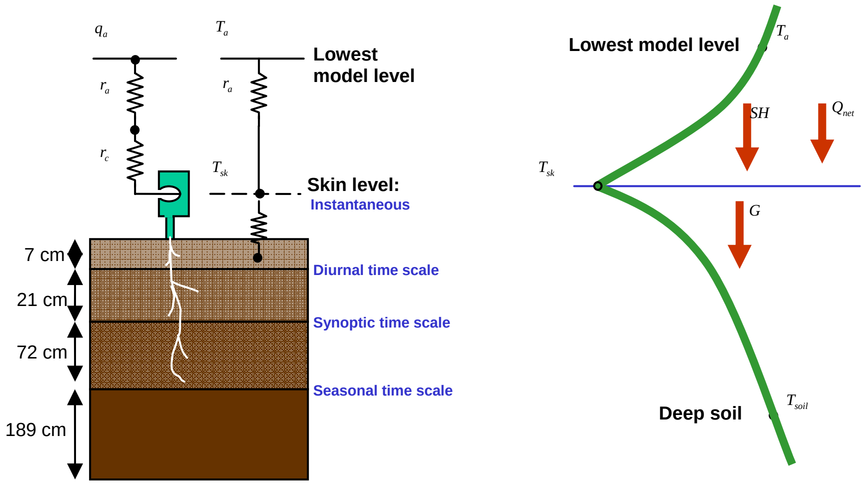


Figure 1 Illustration of the 4-layer land surface model (left; Viterbo and Beljaars 1995) and a schematic temperature profile in soil and lower atmosphere in surface cooling conditions (right).

The skin layer is connected to the first soil layer through an empirical skin layer conductivity. The surface energy balance equation is used as part of the surface boundary condition. The solution for the skin temperature is found as part of the implicit solver for boundary layer diffusion, a procedure that has strong similarities with the Penman-Monteith approach. The soil moisture equation is coupled to the atmosphere through precipitation and evaporation. Evaporation can be from bare soil, from the intercepted rain water or from transpiring vegetation which extracts water from the root zone (up to 1 m deep).

At night or winter at high latitudes, the latent heat flux is small and the skin temperature is mainly controlled by the net radiation at the surface, the strength of thermal coupling with the deeper soil and the strength of the turbulent coupling with the atmosphere. The coupling with deeper soil depends on various parameters like soil thermal properties, skin layer conductivity and whether soil water freezing is included. The coupling with the atmosphere depends on the boundary layer diffusion (Beljaars and Viterbo 1998). In cold conditions when the surface is cooling the atmosphere, the boundary layer is stable, and particularly sensitive to the stable turbulent diffusion which is also uncertain due to heterogeneous terrain effects.

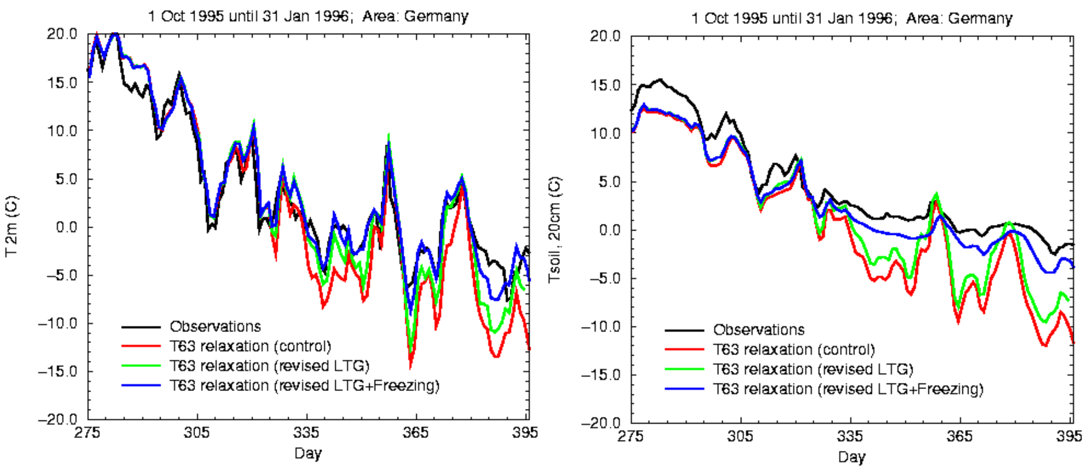


Figure 2 Time series of 2m air temperature (left figure) and soil temperature at a depth of 20 cm (right figure) for the winter of 1995/1996. Observations are shown in black averaged over Germany and compared to simulations with the control model, the model with revised boundary layer diffusion (revised LTG), and a model version with both revised boundary layer diffusion and soil moisture freezing. Model soil temperatures are from layer 2.

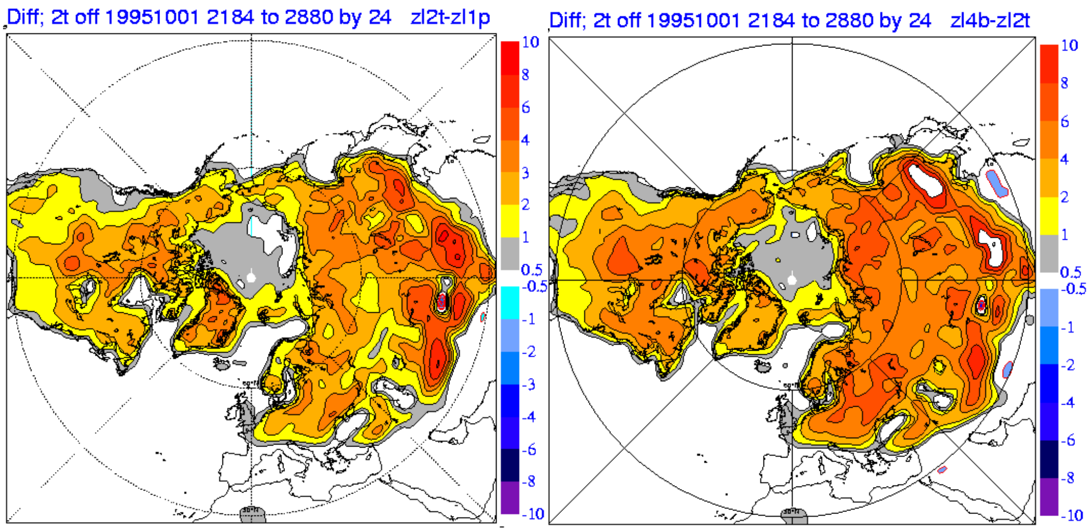


Figure 3 Air temperature difference (at 2m) for January 1996 between a simulation with the revised BL diffusion and the control (left figure), and between the revised BL diffusion+ soil moisture freezing and the control (right figure).

The sensitivity of the 2m temperature (which is related to the skin temperature) to atmospheric and soil coupling is demonstrated here with two model changes that were introduced in the ECMWF model to alleviate winter time temperature biases over continental areas (see Viterbo et al. 1999 for full details). The atmospheric coupling was increased by increasing the diffusion coefficient for heat in the stable boundary layer. The inertia of the land surface was increased by introducing the process of soil moisture freezing which slows down the temperature drop around zero degrees due to the latent heat necessary for the freezing process. The effect of these changes is illustrated in the Figs. 2 and 3, making use of long simulations (starting 1 Oct 1995) in which the atmosphere above 600 m is relaxed towards the analysis. Such simulations have the advantage that the synoptic evolution of a particular year can be reproduced, while the land surface interaction on a seasonal time scale can be studied. Figure 2 shows the 2m temperature (from SYNOP's) and soil temperature averaged over a large number of stations in Germany. The control model has a pronounced cold bias, which was particularly bad in the winter of 1995/1996 due prolonged blocked weather conditions over central Europe. Both the increased boundary layer diffusion and the introduction of soil moisture freezing reduce the biases substantially. As to be expected, the impact of boundary layer diffusion is most visible in 2m temperature, while the effect of soil moisture freezing dominates in the soil temperatures.

Fig. 3 shows the impact from increased boundary layer diffusion and soil moisture freezing on the Northern Hemisphere 2m temperatures in January 1996. The impact of both changes is substantial and of comparable magnitude. This pronounced sensitivity of the near surface temperature to atmospheric and deep soil coupling may explain some of the uncertainty in climate change results, because temperature change in simulations of future climate often shows strong signals over continental areas in winter, while the specification of the relevant coupling coefficients is rather uncertain.

3. Snow albedo in forest areas

Snow has a strong impact on surface albedo, so in a model it is important to have a good representation of snow cover and if snow is present to assign the correct albedo to the snow covered areas. In 1996, the ECMWF model had a snow albedo of 0.8 in all areas with snow cover. During the BOREAS experiment, the albedo was measured extensively over forest and it clearly indicated that the value of 0.8 was not representative because most of the snow is underneath the forest canopy. Observed albedo values for the BOREAS aspen and conifer sites were around 0.2, which is also the value that was introduced at the end of 1996 in the ECMWF model (Betts and Viterbo 1999). The effect is very clear on the heating of the lower troposphere in spring when snow is still present over large areas of the North American and Asian continents and the solar elevation is already high enough for the solar radiation to play a substantial role in the boundary layer heating. Fig. 4 shows the effect on systematic forecast errors. With the old albedo of 0.8, the lack of heating results in a wide spread cold bias. After the reduction of the albedo to 0.2, the systematic errors have been mostly eliminated.

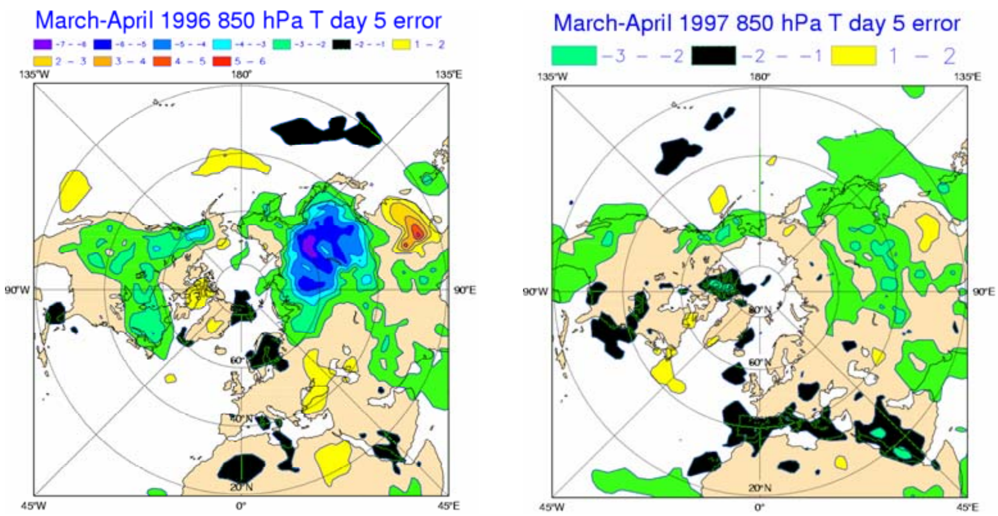


Figure 4 Mean 850 hPa day-5 temperature error (day 5 forecasts minus analysis) from the operational forecasts in April/May 1996 (left figure) and after the snow albedo was reduced for forest areas in April/May 1997.

4. Sea ice temperature

Until June 2000, the ECMWF model used a single ice slab model (thickness 2 m) to describe the temperature evolution of sea ice. The thick ice slab is obviously not capable of representing multiple time scales. Therefore with the introduction of the tiled land surface scheme TESSEL, a 4-layer sea model was included. Making use of the tile structure of TESSEL, ocean grid boxes are subdivided in an open water tile and a sea ice tile. The sea ice fraction and the open water temperature are prescribed from an analysis that is provided daily by NCEP (with sea ice cover from SSMI).

The sea ice model has 4 layers with thicknesses of 7, 21, 72 and 50 cm. The top of the 7 cm layer is used as a skin layer with conductivity between the skin and the middle of the 7 cm layer according to the heat diffusion coefficient of ice. The thickness of the deep layer was optimized using observations of the seasonal evolution of air temperature over sea ice. The boundary condition at the bottom is set at -1.7^{o} C. The albedo of the sea ice is prescribed from the monthly climatology by Ebert and Curry (1993). Melting ponds have big impact on albedo but are not parametrized to avoid the positive feedback between melting, albedo and temperature. Also snow cover on sea ice is not included, because there is no routine data of snow depth on sea ice to constrain this parameter.

To test and optimize the sea ice model, ice buoy data was used from 1998/1999, kindly made available by the arctic research community (thanks to I. Rigor and M. Serreze). So-called relaxation runs were used in which the atmosphere above 600 m from the surface was relaxed to an operational analysis. This has the advantage that the weather of one particular year can be realized in a single long integration. The depth of the 4th layer was adjusted to obtain a realistic seasonal cycle compared to the ice buoy observations. Comparison of these original simulations with ERA-40 daily forecasts (ERA-40 used the 4-layer ice model) show nearly identical results so only ERA-40 results are shown here.

Fig. 5 shows a time series of daily 24-hour forecasts (verifying at 12 UTC) from the operational model in 1998/1999 (which a single ice slab model) and from ERA-40 (with the 4-layer model) compared to observations at one location. It is clear that the new 4-layer model follows the synoptic variability much better than the old slab model; the four layers are necessary to accommodate for a wide range of time scales. However, the variability at the shortest time scales is still underestimated. This is even clearer in Fig. 6, which shows for April/May 1999 the diurnal cycle of air temperature and skin temperature together with the 4 terms of the surface energy balance from ERA-40 in the right hand panel. The amplitude of the diurnal cycle is nearly 10^{o} C in the observations whereas it is only a few degrees in the model. The large diurnal cycle must be limited to a very shallow layer near the surface because the diurnal amplitude of the radiative forcing is only about 50 W/m^2 and the radiative forcing barely reaches positive numbers during day time. The dominant terms in the surface energy balance are the net radiation and the heat flux into the ice, with some imbalance due to sensible heat flux. Without observations, it is difficult to say how realistic these fluxes are.

The lack of variability in temperature and the underestimation of its diurnal cycle are clear model deficiencies. It means that a fast response layer near the surface is missing. Adding snow on top of the ice might improve this because snow acts as thermal insulation between the atmosphere and the underlying ice. However, it may be difficult to control such an additional variable (snow depth) in a data assimilation and forecasting system because no routine observations of snow depth over ice exist.

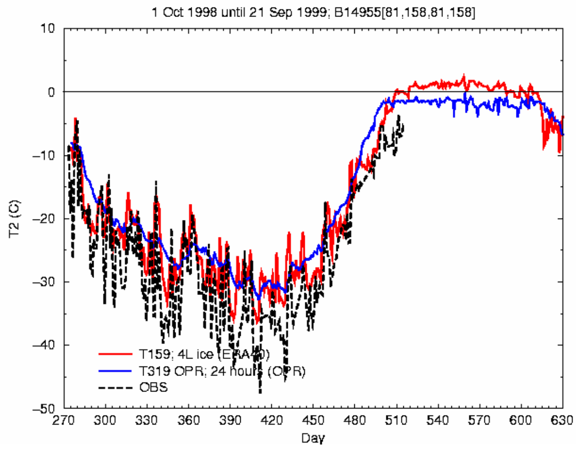


Figure 5 Time series (19981001-19990921) of daily 24-hour forecasts of 2-m temperature from the ECMWF operational model (with the single slab model; blue) and the daily 24 hour forecasts from ERA-40 (with the 4-layer sea ice model, red) compared to ice buoy observations (black) at about 81°N/158°E.

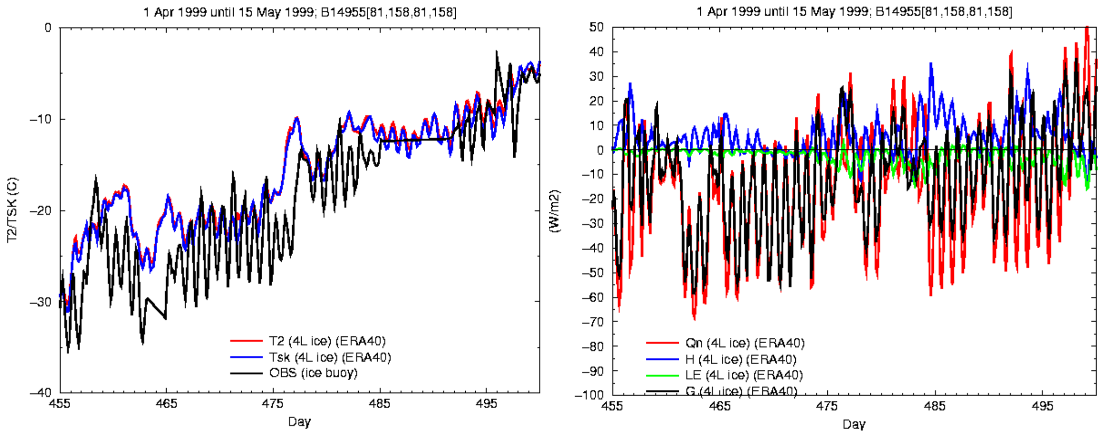


Figure 6 Left panel: Time series (19990401-19990515) of daily 6,12,18,24-hour forecasts of 2-m temperature (red) and skin temperature (blue) from ERA-40, compared to ice buoy data (black). The red and the blue curve are difficult to distinguish because they are nearly identical. Right panel: Time series of the 4 components of the surface energy balance in ERA-40 i.e. net radiation (red), sensible heat flux (blue), latent heat flux (green) and heat flux into the ice (black). Units are W/m^2 and downward fluxes are positive.

5. Snow over land

To have a simple representation of surface heterogeneity, the Tiled ECMWF Scheme for Surface Exchanges over Land (TESSEL) was introduced in 2001 (van den Hurk et al. 2000). To characterise the land use, it makes use of the GLCC global climatology for high vegetation cover, low vegetation cover, high vegetation type and low vegetation type. Seventeen land cover types are distinguished. Over land, TESSEL has 6 sub-areas with variable fraction for each grid point: High vegetation, Low vegetation, Wet Interception layer, Bare soil, Exposed snow and Snow under high vegetation (Fig. 7). The distinction of high and low vegetation type is particularly important for snow, because exposed snow (i.e. low vegetation covered by snow) has a high albedo and snow under high vegetation has little impact on the forest albedo. The tile structure allows the model to account for this in a consistent way.

Snow is represented as a single layer with prognostic variables for snow mass, density and albedo following Douville et al. (1995). For low snow amounts, snow cover is linearly related to snow mass with full cover above 15 kg/m^2. Snow water equivalent is evolved with tendencies from precipitation, melting and evaporation. Snow albedo (for exposed snow) and density use relaxation equations. Snow density is 100 kg/m^3 for fresh snow evolving to 300 kg/m^3 with a rather fast time-scale of 3 days.

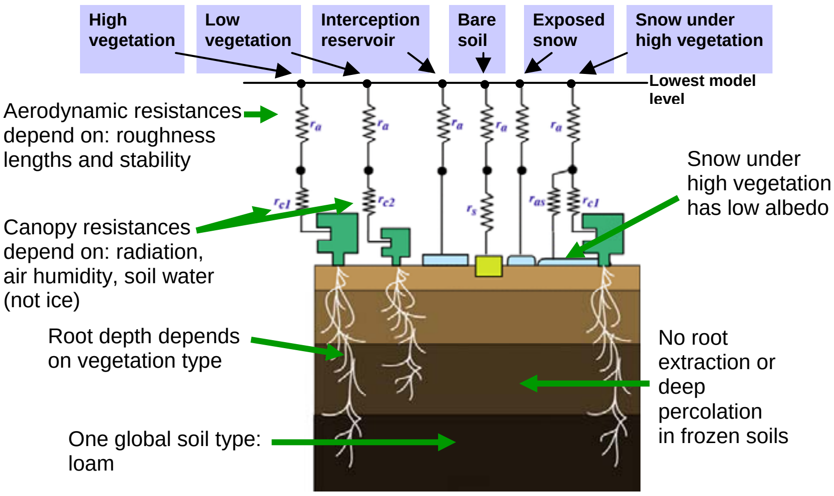


Figure 7 Illustration of the Tiled ECMWF Scheme for Surface Exchanges over Land (TESSEL). The soil layer depths are 7, 21, 72 and 189 cm respectively.

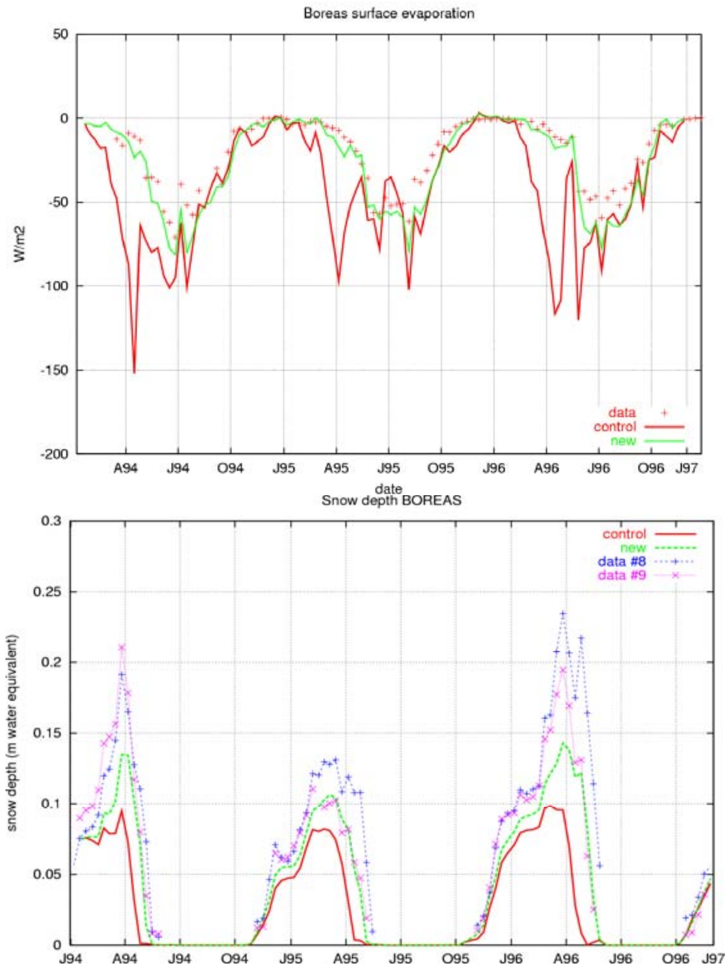


Figure 8 Time series (10-day averages) from January 1994 to January 1997 of Latent heat flux (top panel) and snow depth (bottom panel) for BOREAS. The plusses and crosses are observations; the red curve represents the old model (no tiles); and the green curve represents TESSEL.

Albedo is reset to 0.85 for fresh snow and relaxes to 0.50 with a time scale of a month for cold snow and about 4 days for melting snow.

The temperature of the one-layer snow pack is controlled by the surface energy balance, the heat exchange between snow and first soil layer and the latent heat of melting. To obtain a reasonable amplitude diurnal cycle with the single layer snow model, the snow depth in the temperature equation (as used for the thermal inertia) is assumed to be 7 cm. Melted snow disappears immediately from the snow mass, because refreezing of water in the snow pack is not considered. For exposed snow, the skin temperature is the temperature of the top of the snow layer. For the tile with snow under high vegetation, the vegetation canopy is represented by the skin temperature and the "skin layer conductivity" (which is different for stable and unstable stratification) is the coupling coefficient with the underlying snow deck. This configuration is quite important in spring, because it allows the canopy temperature (skin temperature) and therefore also the air temperature to rise well above 0^oC due to solar heating while there is still melting snow on the ground at 0^oC.

TESSEL was tested extensively in stand alone mode using observational data sets. Fig. 8 shows a simulation with the old (non-tiled) model and with the TESSEL model. A major difference is that TESSEL eliminates the unrealistic spring peak in evaporation. The old model used all the available energy (which is high due to the low forest albedo) for evaporation of snow. The new model has a more realistic coupling between canopy and snow and due to the frozen soil, transpiration is still not possible. Therefore, most of the available radiative energy is transferred through sensible heat flux into the atmosphere and a smaller amount is used for snow melt. The depletion of snow in spring is also slower and in better agreement with observations (bottom panel of Fig. 8).

6. TESSEL in ERA-40

The TESSEL scheme was used as part of the assimilation model for a 45 year long atmospheric and land surface re-analysis from 1957 to 2003 known as ERA-40 (Uppala et al. 2006). Here we focus on a few high latitude aspects: the distribution of permafrost, analysis increment for soil moisture and temperature and comparison with BERMS data.

6.1. Permafrost

Figure 9 shows the distribution of permafrost in ERA-40 and the map published by the International Permafrost Association. Ten-year monthly averages from 1986 to 1995 are used to determine the maximum

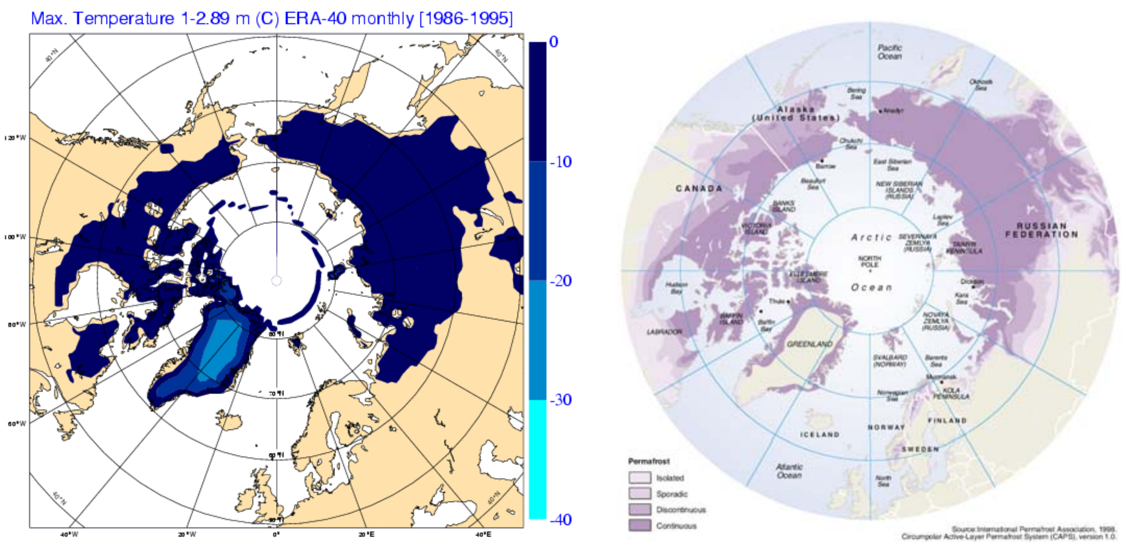


Figure 9 Distribution of permafrost in ERA-40 from 1986 to 1995 (left panel) and from the International Permafrost Association (right panel; light, medium and dark shading indicate sporadic, discontinuous and continuous permafrost respectively). For ERA-40, the maximum temperature of all 4-layers is plotted from 10-year monthly means. Shading starts below 0^oC and is interpreted as permafrost.

temperature of all soil layers. Resulting temperatures below zero are interpreted as permafrost. It can be seen that the main climatological distribution is well reproduced. It means that with a realistic model it is possible to control the soil temperatures by atmospheric data assimilation. This comparison gives information about the model's capability in reproducing the summer/autumn temperature when the deep soil is warmest. In section 7, winter temperatures will be compared to BERMS data.

6.2. Data assimilation increments

Soil temperature and soil moisture are not observed routinely in a global network. Therefore ERA-40 (and the operational ECMWF system) uses SYNOP observations to infer soil temperature and soil moisture (Douville et al. 2000). Snow depth is analysed directly from SYNOP observations. The idea of using atmospheric temperature and moisture observations to infer soil parameters is that with a good atmospheric model, a drift in temperature and/or moisture will occur in short range forecasts (6 to 12 hours) if the surface fluxes are biased. In unstable situations, the drift is interpreted as an error in soil moisture. In stable conditions the drift is attributed to soil temperature. Snow mass is analysed from SYNOP snow depth observations, with an additional nudging towards climatology (12-day time scale). The latter is important in areas without any observations. Also the use of SYNOP snow depth observations is not without problems. Because snow density is not observed, snow depth is converted to snow mass using the model snow density (which might be in error).

In a well balanced system, data assimilation corrects for random errors only. If systematic increments occur, it means that the underlying model is biased. So it is worth looking at mean data assimilation increments as indication of possible model errors. Interpretation is by no means trivial, because it is not always clear which model aspect is in error. It is for instance not guaranteed that systematic soil moisture/temperature increments are only due to the land surface scheme. The assumption of an unbiased atmospheric model may also be wrong, e.g. errors in boundary layer mixing may lead to boundary drift in the short range forecasts.

Figure 10 shows the data assimilation increments of 2m temperature, relative humidity and snow water equivalent (left panels), which are due to observations at SYNOP stations (and snow climatology) and the resulting soil temperature and soil water increments (right panels). The discussion in this paper is limited to the NH extratropics. The temperature increments in winter are predominantly negative with a clear correspondence between air temperature and soil temperature. The negative increments are the consequence of the model tendency of being too warm in stable situations. The soil data assimilation compensates for this by lowering the soil temperature in the top soil later. Inspection of the geographical distribution of these increments in January shows that increments are typically of the order of 1 K over the N-American and Asian forest areas. Such increments, which occur every 6-hourly analysis cycle, are by no means small. A 1 K temperature change in 6 hours in a 7 cm soil layer with the model heat capacity of 2.2 10^6 J $K^{-1}m^{-3}$, corresponds to a heat flux of 7.2 W/m^2, which is a substantial energy input into the land surface if applied systematically over a long period of time.

Also RH increments show a systematic seasonal pattern with drying in spring (the band with negative increments moving from 40^oN to 70-80^oN in June) and moistening in summer. This pattern is mirrored in the soil water increments. It is believed that the seasonal evolution of increments is related to a too small water holding capacity of TESSEL. This was documented by Hirshi et al. (2006) who analysed terrestrial water evolution in river catchments using atmospheric moisture convergence from ERA-40 and observed run-off. The seasonal cycle of terrestrial water in ERA-40 has only half the amplitude of that observed, because the data assimilation increments damp the seasonal cycle. A new hydrology formulation with a geographical distribution of soil properties is currently under development and will increase the soil water holding capacity.

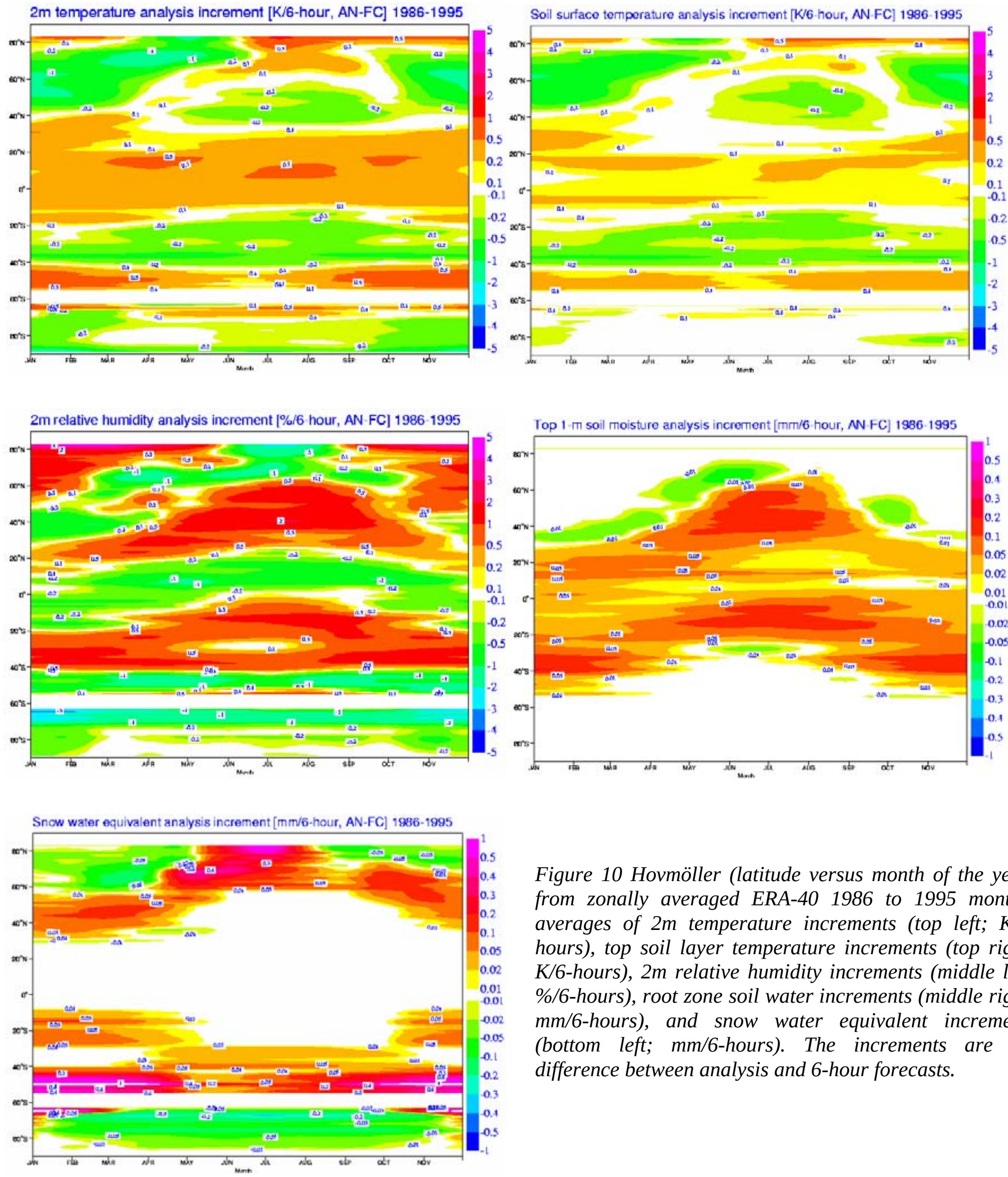


Figure 10 Hovmöller (latitude versus month of the year) from zonally averaged ERA-40 1986 to 1995 monthly averages of 2m temperature increments (top left; K/6-hours), top soil layer temperature increments (top right; K/6-hours), 2m relative humidity increments (middle left; %/6-hours), root zone soil water increments (middle right; mm/6-hours), and snow water equivalent increments (bottom left; mm/6-hours). The increments are the difference between analysis and 6-hour forecasts.

Increments in snow mass are also substantial with negative increments in winter and positive increments in spring. The spring signal is consistent with the results of Betts et al. (2003) for the Mackenzie basin. However, for winter they find positive increments which they attribute to the use of too high snow density in cold temperatures (the model relaxes towards 300 kg/m^3 in 3 days) and with density overestimated, the snow depth observations lead to positive snow mass increments. The positive increments in spring suggest a too quick snow melt in the model. This may be due to the small snow depth of 7 cm in the snow temperature equation. With the single snow layer, the temperature might reach 0^oC too frequently and then loose melt water too quickly as there is no provision for refreezing.

As was shown before, TESSEL was a clear improvement over the non-tiled model. However, inspection of Fig. 8 with the single column BOREAS simulation shows that TESSEL still melts snow too quickly. We will come back to this question in the next section using BERMS data.

6.3. Comparison to BERMS data

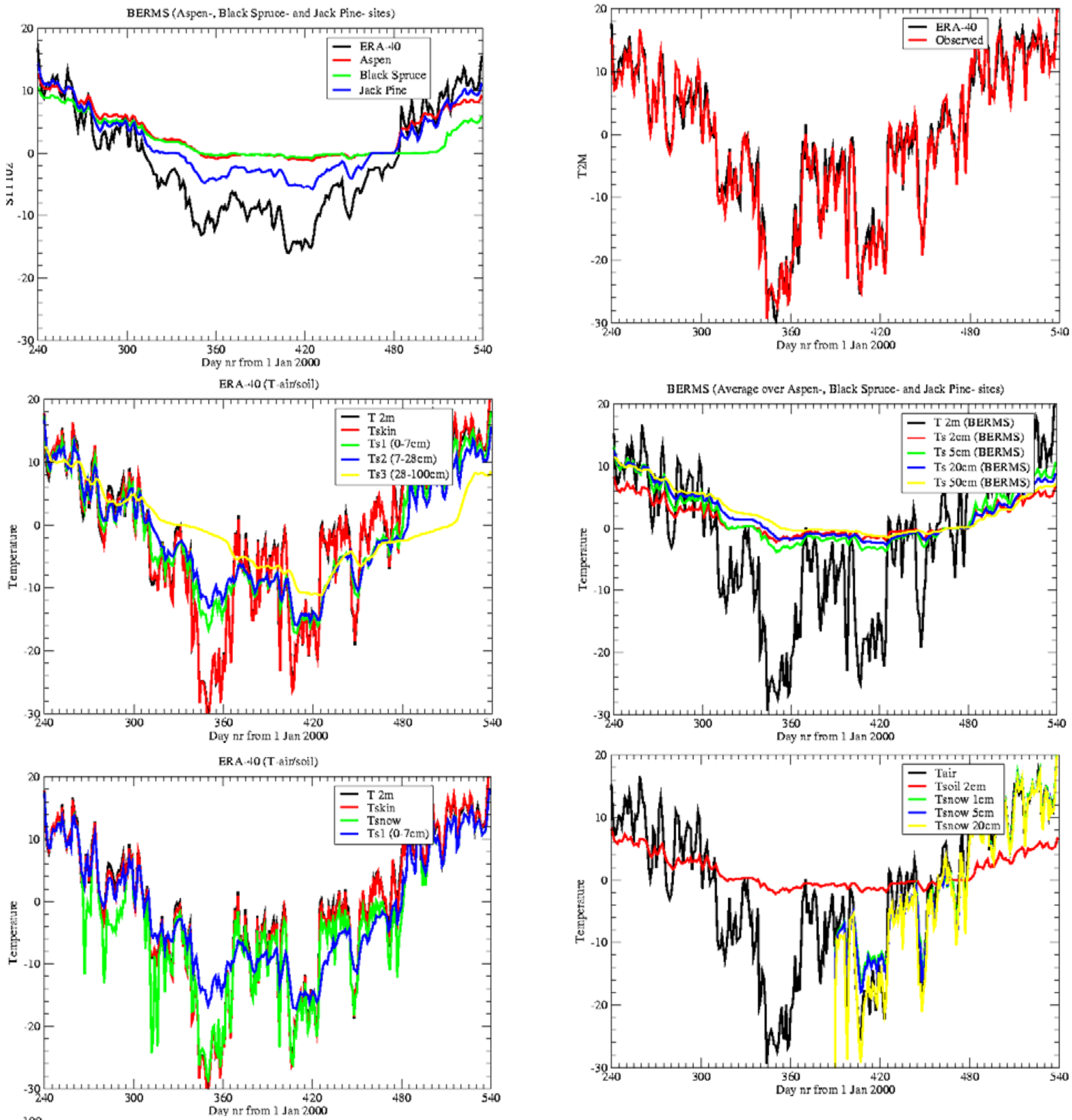
As shown in section 5, off-line evaluation using observational data is an important step in the development phase of a land surface scheme. The Boreal Ecosystem Atmospheric Study (BOREAS) has played an important role in the development of land surface schemes at ECMWF (Betts et al. 2001). The facilities of this experiment were expanded and converted into a more continuous monitoring project through the Boreal Ecosystem Research and Monitoring Sites (BERMS) project. The BERMS sites have been operating for many years and provide high quality data that can be used for model evaluation and parameter optimization. Here we use the BERMS data to look at the thermal coupling of boreal forest to the atmosphere (for a more comprehensive study see Betts et al. 2006).

BERMS samples the land surface heterogeneity with observations from 3 contrasting sites less than 100 km apart in Saskatchewan at the southern edge of the Canadian boreal forest (at about 54^{o}N/105^{o}W). The three locations are: (i) the Old Aspen site (deciduous, open canopy, hazel under-story, 1/3 of evaporation from under-story), (ii) the Old Black Spruce site (boggy, moss under-story), and (iii) the Old Jack Pine site (sandy soil). The ERA-40 model has a resolution of about 125 km and therefore all sites are basically part of a single grid box. The closest grid point in ERA-40 has 98% cover with evergreen needle leaf forest for the high vegetation tile and 2% cropland / mixed farmland for the low vegetation tile.

The first question that arises is whether the real heterogeneity of the terrain can be represented by the simple tile structure that only distinguishes between high and low vegetation. The model has 98% high vegetation with a single vegetation type and the tile structure is only relevant in relation to snow cover and intercepted water. To see how heterogeneous the terrain is in its effect on thermal coupling, the soil temperature (20 cm deep) is plotted in comparison to the ERA-40 temperature in layer 2 (Fig. 11, top left). All the time series are 24-hour hour averages. The ERA-40 time series are also 24-hour averages, but from daily 0-24 hour forecasts starting from the analysis at 0 UTC. The Aspen and Black Spruce sites show a very similar evolution of the soil temperature with hardly any drop below freezing. The Jack Pine site is different from the other two in the sense that the soil temperature shows more variability and that temperature drops a few degrees below zero in cold spells. Apparently the thermal coupling between atmosphere and soil is very weak for the Aspen and Black Spruce sites and slightly stronger for the Jack Pine site. ERA-40 shows a lot more variability and the temperature drops to about -15 ^{o}C during the coldest phase of winter. So the thermal coupling is a lot stronger in ERA-40 than in the real world.

Because the differences in thermal coupling is much smaller between sites than between observations and ERA-40, in the remaining figures the three sites are averaged before comparing with ERA-40. Ideally, the averaging should have been weighted with the relevant vegetation type fractions in the ERA-40 grid box, but these fractions were not available, so a simple average is taken.

The 2m temperature is compared in the top right panel of Fig. 11. It shows that ERA-40 follows the synoptic and seasonal variability quite accurately. These results are consistent with those of Betts et al. (2006; see also Simmons et al. 2004 for a global perspective). The observation that the soil temperature drops barely below freezing is the more remarkable given the very cold air temperatures that are reached sometimes. To see how the temperature variability penetrates into the soil, the 2m temperature and the soil temperatures are shown at various levels in the second row of Fig. 11 with ERA-40 on the left and observations on the right. For ERA-40 also the skin temperature is shown, but it is virtually identical to the 2m temperature for these daily means. The observations show a big jump in variability when going from the atmosphere into the soil, whereas ERA-40 shows a much more gradual decrease of variability. The atmospheric temperature signal penetrates much deeper into the soil in ERA-40 than observed. This raises the question where the jump



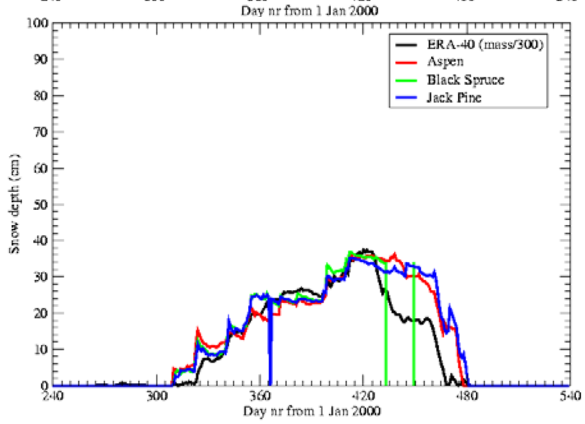


Figure 11 Time series of diurnal averages from 27 August 2000 (day 240) to 23 June 2001 (day 540) of observed and daily 0 to 24 hour ERA-40 forecasts from 0 UTC. The model data have been averaged from hourly output for the BERMS location. The top left figure compares soil temperature of layer-2 of ERA-40 with the observed soil temperature 20 cm deep at the Aspen, Black Spruce and Jack Pine sites of BERMS. For all the other figures (except the lowest panel) the 3 sites are averaged. The top right figure compares BERMS and ERA-40 2m temperature. The second row of figures shows temperature at 2m and at different depths in the soil with ERA-40 at the left and BERMS at the right. The third row shows the air, soil and snow temperatures at different levels (in BERMS the height in the snow pack is measured from the surface). The lowest panel shows the snow depth in ERA-40 and observed.

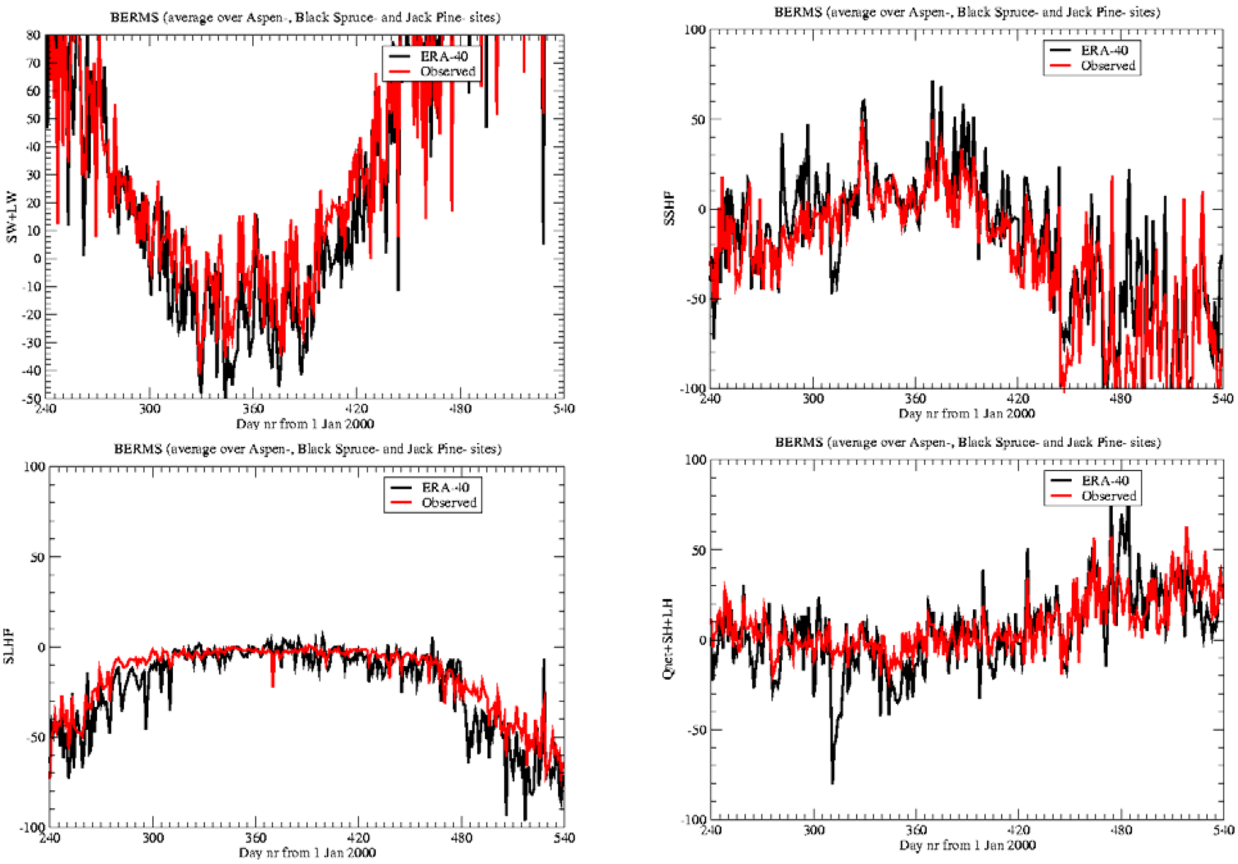


Figure 12 Time series of the diurnally averaged surface energy balance from 27 August 2000 (day 240) to 23 June 2001 (day 540) from observations and daily 0 to 24 hour ERA-40 forecasts from 0 UTC. The top left figure compares net radiation, the top right surface sensible heat flux, the bottom left latent heat flux and the bottom right show the residual of the 3 others i.e. the energy that goes into the land surface.

actually occurs. Therefore the temperature of the snow underneath the forest canopy is considered in row 3 of Fig. 11. ERA-40 has only one snow temperature and the observations have various levels (the height in the observations is measured from the soil surface). The main conclusion is that in ERA-40 as well as the observations, the daily mean snow temperatures follow the atmospheric temperatures, and that these temperatures are highly decoupled from the soil temperatures in the observations. The model has much more coupling between the snow and the soil temperature than observed. The explanation is not obvious, but there must be an insulation layer between the snow and the soil which is not included in the ERA-40 model. The undergrowth, the moss and the litter on the forest floor probably keep a lot of air trapped below the snow deck providing a good insulation between the snow and the soil. However, it is worth noting that the problem of excessive coupling in ERA-40 is not limited to snow conditions. Even before snow is present (see bottom panel of Fig. 11), the coupling appears to be too strong, which reinforces the hypothesis of a litter layer insulating the ground.

The last panel of Fig. 11 also illustrates that the snow disappears too quickly in ERA-40, which is consistent with the analysis increments that were shown in section 6.2. As soon as the air temperature reaches $0^{o}C$ e.g. around day 430 and day 460, snow melts very quickly in ERA-40, much faster than indicated by observations. The reason is not entirely clear, but somehow too much heat must go into the snow layer. Therefore it is very likely that the coupling between the canopy and the snow layer is too strong. This coupling is represented by the so-called “skin layer conductivity” in the ERA-40 model. For the snow under vegetation tile it is set to 15 $Wm^{-2}K^{-1}$ for stable stratification (i.e. canopy warmer than snow) and to 20 $Wm^{-2}K^{-1}$ for unstable stratification. Also the representation of the thermal inertia of the snow by a 7 cm layer may result in a too frequent temperature rise above $0^{o}C$ resulting in spurious melting.

In Fig. 12, the modelled and observed surface energy components are considered, namely net radiation, sensible heat flux, latent heat flux and ground heat flux. Although the ground heat flux is measured directly (with heat flux plates in the soil), it is not used here, because it would exclude the energy that is absorbed by the canopy and the snow. So the ground heat flux is defined as the residual of atmospheric fluxes i.e. the sum of net radiation, sensible and latent heat flux. There are no obvious systematic biases in the radiative and turbulent fluxes except a slight overestimation of latent heat flux in autumn and spring. Even in the residual ground heat flux, the correspondence between observations and model is quite good. However, it should be realized that the fast snow melt is related to small errors in energy flux. Just after day 420 ERA-40 looses about 10 cm of snow in 8 days. To melt this amount of snow in 8 days a mean energy flux of about 15 Wm^{-2} is needed (assuming a density of 300 kgm^{-2}). This level of precision is probably not achievable as the residual of the three observed energy fluxes.

7. Conclusions

A short overview has been given of model developments that led to the Tiled ECMWF Scheme for Surface Exchanges over Land (TESSEL). TESSEL was also used as part of the assimilating model for the ERA-40 reanalysis, which provides a wide range of options for verification. The use of observations from long field experiments at various sites has played a major role during the development, evaluation and improvement of the land surface scheme at ECMWF. The near surface air temperatures in winter turn out to be sensitive to thermal coupling of the land surface to atmosphere and to the deeper soil. This sensitivity is illustrated with two model changes, namely a change in the stable boundary layer diffusion and the introduction of soil moisture freezing. Both changes had a big impact on the 2m temperature simulations in winter.

At high latitude, albedo has a strong impact on the boundary layer heating. In spring the albedo is still affected by snow, whereas the solar heating is already sufficient to be affected by albedo. From the BOREAS experiment it became clear that the presence of snow under forest increased the albedo rather little because the canopy remains dark even with snow underneath. Changing from a high albedo in forest areas with snow to the more realistic low one reduced the spring time temperature errors substantially.

Sea ice is also handled by the tile scheme to represent ocean points with partial ice cover. The variability of the atmospheric near surface temperature from diurnal to annual time scale is influenced by the response time of the sea ice boundary condition. It is demonstrated that more layers allow for a better representation of various time scales than a single slab model. TESSEL uses 4 layers for ice, but the temperature variability is still underestimated. The amplitude of the diurnal cycle of temperature in spring is underestimated by a factor 3. A snow layer on top of the ice is probably necessary to cure this problem.

The handling of snow is an important aspect of a land surface scheme at high latitudes. It has been demonstrated that the tile structure in TESSEL is particularly relevant because exposed snow and snow under high vegetation are represented in different tiles. This led to an improvement with respect to the non-tiled scheme in which snow was evaporated rather quickly in spring.

Data assimilation increments in ERA-40 have been discussed, because they give information about model deficiencies. The negative increments in spring together with the positive increments in summer suggest that the water holding capacity of the soil is too small. A better representation of the land surface hydrology and the soil properties is expected to improve this. ERA-40 also shows systematic increments of snow depth in spring which indicates that the model snow melt is too fast. The reason is not clear, but the melting process is sensitive to the aerodynamic coupling to the atmosphere through the boundary layer scheme and also to the sub-canopy aerodynamic coupling.

Comparison with BERMS data shows that ERA-40 model / data assimilation system is very efficient in controlling the 2m temperature to a fairly high level of accuracy. However, the comparison also indicates that the thermal coupling of the atmosphere to the soil is generally too strong in the model. The observations show that the there is a pronounced decoupling between the snow and the soil which suggest that an insulating layer exists below the snow e.g. due to trapped air in the undergrowth. Such an effect is not represented in the model. Future model development will be oriented towards less coupling between the atmosphere and the soil e.g. through a less diffusive stable boundary layer scheme and through adjustments in coefficients in the land surface scheme. However, this will reduce the control over boundary layer temperatures through the land surface scheme and its data assimilation, so it may deteriorate the system unless all components are well balanced, i.e. unbiased clouds radiation, precipitation etc.

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